PORTLAND HARBOR RI/FS

APPENDIX LA
SEDIMENT TRANSPORT MODELING
DRAFT FEASIBILITY STUDY

DRAFT

DO NOT QUOTE OR CITE
This document is currently under review by US EPA and its federal, state, and tribal partners, and is subject to change in whole or in part.

March 30, 2012

Prepared for
The Lower Willamette Group

Prepared by
Anchor QEA, LLC

TABLE OF CONTENTS

LIST OF FIGURES

LIST OF TABLES

LIST OF ACRONYMS

µm	micrometer
2D	Two-dimensional
3D	Three-dimensional
ADCP	Acoustic Doppler Current Profiler
AEP	Annual Exceedance Probability
cfs	cubic feet per second
cm	centimeter
cm/yr	centimeter per year
DEA	David Evans and Associates
EFDC	Environmental Fluid Dynamics Code
EPA	U.S. Environmental Protection Agency
FS	Feasibility Study
g	gram
GSD	Grain Size Distribution
IACWD	Interagency Advisory Committee on Water Data
m	meter
m/day	meters per day
mg/L	milligrams per liter
MT/yr	Metric Tons per year
MVUE	Minimum Variance Unbiased Estimator
NOAA	National Oceanic and Atmospheric Administration
NSR	Net Sedimentation Rate
Pa	Pascal
RM	River Mile
Site	Portland Harbor Superfund Site
TOC	Total Organic Carbon
TSS	Total Suspended Sediment
USGS	U.S. Geological Survey
WSE	Water Surface Elevation

EXECUTIVE SUMMARY

Appendix La presents the development and calibration of hydrodynamic and sediment transport models of the Lower Willamette River. These models provide transport and sediment flux information that are used as inputs for the chemical fate and transport model presented in Appendix Ha. The model domain includes the Lower Willamette River, Multnomah Channel, and a portion of the Columbia River, with the primary focus being on the Site, from river mile [RM] 1.9 to RM 11.8. A large of amount of Site-specific data was used to define system geometry, initial conditions, boundary conditions, bed properties and parameters used as input to the hydrodynamic and sediment transport models. Both of the models were successfully calibrated, with the sediment transport model using bed elevation change data collected during a multi-year period to evaluate model performance over a wide range of spatial scales. A sensitivity analysis was conducted to evaluate the effects of uncertainty in various model inputs on the predictive capability of the sediment transport model. The calibrated model was used to simulate a high-flow event (i.e., January 1996 flood) and a 45-year period.

1.0 INTRODUCTION

1.1 STUDY OBJECTIVES

This report details the development of a hydrodynamic and sediment transport model of the Lower Willamette River downstream of River Mile (RM) 13 to the confluence with the Columbia River. The model was developed to support the draft Feasibility Study (FS) for the Portland Harbor Superfund Site (Site). The sediment transport model results were used as follows:

- As inputs for the contaminant fate and transport model of the system, which in turn was used to evaluate the current situation and potential remedial alternatives detailed in the main section of this report including the propensity for natural recovery of the system
- To understand the draft FS level designs for cap armoring to protect against current driven erosion (Appendix Hc)
- To evaluate the potential for erosion of buried contamination (see main text, Section 5.7)

1.2 OVERVIEW OF TECHNICAL APPROACH

The original version of the Lower Willamette River hydrodynamic model, developed and calibrated by WEST Consultants and TetraTech between 2004 and 2009, was used as a base for the hydrodynamic model presented in this report. Descriptions of the original hydrodynamic model are presented in WEST Consultants (2004, 2005a, 2005b, 2006) and WEST Consultants and TetraTech (2009). Some of the input parameters for the hydrodynamic model were refined, and the model was re-calibrated during this phase of the modeling study. The sediment transport model is based on the same framework as the hydrodynamic model (i.e., Environmental Fluid Dynamics Code [EFDC]) but was modified by Anchor QEA to incorporate the SEDZLJ algorithm for transport of cohesive and non-cohesive sediment. SEDZLJ is a sediment transport algorithm developed by Anchor QEA personnel over the course of more than 20 years of modeling practice.

The sediment transport model was calibrated using bathymetric data collected within the Study Area (RM 1.9 to 11.8) over a 7-year period between 2002 and 2008. The calibrated model was used to simulate high-flow events to analyze scour potential and also evaluate the potential for natural recovery over multi-year periods (i.e., 45 years). Uncertainty in the model results was analyzed through application of a sensitivity analysis.

This appendix is organized as follows. Section 2 includes a description of refinements to the hydrodynamic model, sediment transport model theory, and a description of model inputs and boundary conditions. In addition, it contains the model calibration procedure and a sensitivity analysis. Section 3 presents an analysis of bed stability during the high-flow event that occurred during the winter of 1996. This section also includes the results of a 45-year simulation. The results of the multi-year simulation were used to analyze the potential for natural recovery within different sections of the river, as described in Appendix Ha.

2.0 MODEL DEVELOPMENT AND CALIBRATION

2.1 GENERAL DESCRIPTION OF MODELING FRAMEWORK

The mathematical modeling framework that was applied to the Lower Willamette River consists of hydrodynamic and sediment transport models that are linked together. The hydrodynamic model accounts for the effects of the following factors on water movement in the Lower Willamette River: freshwater inflow from the surrounding watershed, flow in the Columbia River, and tides. The hydrodynamic model is used to simulate temporal and spatial changes in water depth, current velocity, and bed shear stress. This information is transferred from the hydrodynamic model to the sediment transport model, where it is used to simulate the erosion, deposition, and transport of sediment in the Lower Willamette River. The sediment transport model is used to simulate temporal and spatial changes in suspended sediment concentrations in the water column, bed elevation changes (i.e., bed scour depth and net sedimentation rate [NSR]), and changes in sediment bed composition (i.e., relative amounts of clay, silt, and sand from different sources).

The modeling framework provides a deterministic approach for simulating sediment transport within the Lower Willamette River. The sediment transport model simulates the movement of sediment by suspended load (i.e., primarily clay, silt, and fine sand). Bedload transport of sediment (i.e., near-bed movement of coarse sand and gravel) is not simulated in this study because this mode of sediment transport is minimal within the Lower Willamette River. About 80 percent of the bed area in the Study Area is comprised of a cohesive (muddy) bed, and bedload transport is negligible within a cohesive bed. In addition, no formulations are available from the peer-reviewed literature for simulating bedload transport over a cohesive bed. Within the non-cohesive bed areas (i.e., about 20 percent of the Study Area), successful calibration of the model using bed elevation data demonstrates that the assumption of minimal bedload transport is valid.

The hydrodynamic and sediment transport models are constrained by governing equations that are based on the conservation of mass and momentum. Mechanistic formulations and algorithms are used in the sediment transport model to simulate deposition and erosion of cohesive and non-cohesive sediment. The formulations and algorithms used to simulate deposition and erosion are based on empirical information and data from a wide range of laboratory and field studies (see Section 2.3.2 for a discussion of the theory and equations used in the sediment transport model). In addition, site-specific data are used to determine various parameters used in the sediment transport model, which provides additional constraints on the model.

2.2 DEVELOPMENT AND CALIBRATION OF HYDRODYNAMIC MODEL

2.2.1 General Description

The hydrodynamic model that was applied in this study is EFDC, which is supported by the U.S. Environmental Protection Agency (EPA). EFDC is a three-dimensional (3D) hydrodynamic model capable of simulating time-variable flow in rivers, lakes, reservoirs, estuaries, and coastal areas. For this study, the two-dimensional (2D), depth-averaged hydrodynamic model within EFDC was used. The Lower Willamette River is a freshwater river so only vertical gradients in temperature can generate density-driven circulation. Continuous temperature measurements in 2001 and 2002 (Berger et al. 2004) show minimal vertical temperature gradients, which indicates that density-driven circulation in the river is negligible. Thus, use of a 2D, depth-averaged model is valid for the Lower Willamette River because the effects of density-driven circulation are minimal.

EFDC solves the conservation of mass and momentum equations, which are the fundamental equations governing the movement of water in a freshwater tidal river. The model has been applied to a wide range of environmental studies in large number of rivers, estuaries, and coastal ocean areas. A complete description of the model is given in Hamrick (1992).

2.2.2 Numerical Grid and Geometry

A curvilinear, boundary-fitted numerical grid was used to represent the geometry of the Study Area, with a total of 3,355 grid cells. Figure 2-1 presents the grid extent and the location of the Study Area and the boundary conditions. Cross-sectional variations in bathymetry within the Lower Willamette River are delineated using a range of 5 to 43 grid cells in the cross-channel direction depending on the river width. The upstream boundary of the numerical grid in the Lower Willamette River is located at RM 24. The downstream boundary of the numerical grid extends into the Columbia River so that hydrodynamic interactions between Columbia River and Lower Willamette River are realistically simulated. A one-dimensional grid (i.e., one grid cell-wide) is used in the Multnomah Channel, which extends to the confluence with the Columbia River near St. Helens, Oregon. Spatial variations in bathymetry (i.e., sediment bed elevation) within the Lower Willamette River were specified for input to the model using bathymetry collected during January 2002 (DEA 2002). Figure 2-2 presents the bathymetry projected on the model grid for the Study Area.

2.2.3 Hydrodynamic Model Inputs: Boundary Conditions and Bottom Roughness

The hydrodynamic model requires specification of the following time-variable boundary conditions: 1) inflow at upstream boundary in the Lower Willamette River; 2) inflow at upstream boundary in the Columbia River; 3) water surface elevation (WSE) at downstream boundary in the Columbia River; and 4) WSE at downstream boundary of the Multnomah Channel. Daily-average flow rate data collected at the U.S. Geological Survey (USGS) Portland gauging station were used to specify the inflow at the upstream boundary in the Lower Willamette River for the calibration and long-term simulations. Inflows at the upstream boundary during high-flow events were specified based on the results of a flood frequency analysis. A Log-Pearson Type 3 flood frequency analysis (Helsel and Hirsch 2002) of peak flow rate data from the 36-year historical record was conducted. The Log-Pearson Type 3 analysis is the recommended technique for flood frequency analysis (IACWD 1982). The analysis was conducted as follows:

- Instantaneous peak flow data collected between 1972 and 2007 were used.
- The logarithm of each peak discharge during the 36-year period was calculated.
- For the logarithms of peak discharge, these statistics were calculated: mean (M), standard deviation (S), and skewness (g).
- The peak discharge (Q_p) for a specific annual exceedance probability (AEP) was calculated using:

$$\log(Q_p) = M + KS \tag{2-1}$$

Where,

$\log(Q_p)$ = the logarithm of the peak discharge with an AEP of one in Y years and K is the frequency factor for a specific AEP as a function of skewness (g).

Tabulated values of K are presented in Helsel and Hisrch (2002). For example, AEP values of 10 percent and 1 percent correspond to high-flow events with return periods of 10 and 100 years, respectively. The analysis indicates that a 100-year flood has a flow rate of approximately 360,000 cubic feet per second (cfs). A summary of the estimated flow rates for high-flow events is presented in Table 2-1. For comparison, the annual average flow rate is 33,200 cfs.

Table 2-1. Estimated Lower Willamette River Flow Rates for High-flow Events

Flood Return Period (Years)	Flow Rate (cfs)
2	156,000
10	252,000
25	297,000
50	329,000
100	360,000
500	428,000

Notes:
cfs = cubic feet per second

The original hydrodynamic model specified time-variable WSE at the upstream and downstream boundaries in the Columbia River (WEST Consultants and TetraTech 2009). Initial evaluation of the original model indicated that this approach produced unrealistic circulation patterns in the Columbia River (Figure 2-3). This problem was resolved by specifying flow rate and WSE at the upstream and downstream boundaries in the Columbia River, respectively (Figure 2-4). The closest flow gauge at the upstream boundary of the Columbia River was the USGS gauge at Vancouver, Washington. However, flow rate data at this station were only available for the 4-year period between

October 1965 and September 1969. The next closest USGS gauging station along the Columbia River is located 80 miles upstream at The Dalles, Oregon, where data have been collected since 1878. A correlation between measured flow rates at Vancouver and The Dalles was used to demonstrate that the flow rate at Vancouver can be estimated by adding 13,200 cfs to the flow rate at The Dalles. Thus, a long-term flow record at Vancouver was constructed using the flow rate data collected at The Dalles.

The specification of WSE at the downstream boundary in the Columbia River, and also in the Multnomah Channel, was not changed from the original hydrodynamic model. This boundary was defined using data collected at the National Oceanic and Atmospheric Administration (NOAA) tidal gauging station in St. Helens, Oregon. Boundary condition information (i.e., flow rates in Lower Willamette and Columbia Rivers, WSE at St. Helens) used for the 7-year period (2002 through 2008) to calibrate the sediment transport model is shown on Figure 2-5.

Model input files provided by WEST Consultants indicated that the original version of the model used spatially variable effective bed roughness (i.e., Z_0), as shown in Figure 2-6. In the original version of the model, the effective bed roughness was dependent on Manning's roughness coefficient and water depth using the following relationship (based on Equation 4.1 in WEST Consultants 2006):

$$Z_0 = \frac{H}{2.0\exp\left(\frac{\kappa}{\sqrt{gn^2/H^{1/3}}}\right)} \tag{2-2}$$

Where,

Z_0 = effective bed roughness
H = water depth
n = Manning's roughness coefficient
κ = von Karman's constant.

An evaluation of hydrodynamic model predictions based on spatially variable of Z_0, as calculated using Equation 2-2, indicated that this methodology yielded unrealistic results. Thus, the original model was revised and a spatially constant effective bed roughness was used in the hydrodynamic model. This method is consistent with the approach used in many other hydrodynamic modeling studies (e.g., Blumberg and Mellor 1987). The effective bed roughness (Z_0) was adjusted during calibration of the hydrodynamic model, see Section 2.2.4.

2.2.4 Hydrodynamic Model Calibration Approach and Results

Calibration of the hydrodynamic model was achieved using data collected with an Acoustic Doppler Current Profiler (ADCP) in the main channel of the Lower Willamette River between RM 1 and 11. The ADCP data consisted of measurements of water depth, depth-averaged current velocity (magnitude and direction) during three different periods between 2002 and 2004. A summary of the three ADCP deployment periods is provided in Table 2-2. Two of the survey periods in 2002 and 2003 were conducted at or above mean flow rate (26,000 to 66,000 cfs). The survey conducted in January 2004 was conducted during an approximate 2-year flood event. Sixteen locations were surveyed between RM 1 and RM 11 during the 2002 and 2004 periods. In 2003, only four transects were surveyed near the confluence with the Multnomah Channel. At each location, the survey was made perpendicular to the river flow.

Table 2-2. ADCP Data Collection Summary

Survey Date	Lower Willamette River Flow Rate (cfs)	Survey Region	Number of Transects
April 19, 2002	66,000	RM 1 – 11	16
May 13, 2003	26,000	RM 2.5 – 4	4
January 31, 2004	139,000	RM 1 – 11	16

Notes:
cfs = cubic feet per second
RM = river mile

The model parameter that was adjusted to achieve the optimum agreement between predicted and observed water depth and current velocity was the effective bed roughness (Z_0) in the hydrodynamic model, which represents the total roughness due to form drag and skin friction. Generally, Z_0 ranges from about 0.1 to 10 centimeters (cm). A value of 1 cm for effective bed roughness produced the best agreement between observed and predicted water depth and depth-averaged current velocity during the calibration period. Figures 2-7 through 2-12 show examples of the model-data comparisons at three locations: RM 10, RM 6, and RM 3. These figures show that the model adequately predicts current velocity over a wide range of flow rates. A comparison of predicted and measured WSE at the USGS Portland gauging station during 2008 is presented in Figure 2-13. Model predictions of WSE are in good agreement with measured values during low-flow condition, but the model tends to underpredict WSE during high-flow conditions. The underprediction during high-flow conditions is primarily due to specification of WSE at the downstream boundary using the data collected at the St. Helens gauging station. The underprediction of WSE during high-flow conditions, however, will tend to produce conservative results because it will cause an overprediction of bed shear stress, which will produces higher erosion and lower deposition predicted by the sediment transport model. Overall, the calibration results indicate that the hydrodynamic model is able to adequately simulate WSE, water depth and current velocity in the Lower Willamette River for a wide range of river flow and tidal conditions.

2.3 DEVELOPMENT AND CALIBRATION OF SEDIMENT TRANSPORT MODEL

2.3.1 Overview of Sediment Transport Model Structure and Capabilities

The sediment transport model used in this study, referred to as SEDZLJ, is capable of simulating erosion and deposition of sediment within cohesive (i.e., muddy) and non-cohesive (i.e., sandy) bed areas (Ziegler et al. 2000; Jones and Lick 2001; QEA 2008). The sediment transport model has the following characteristics and capabilities: 1) transport of suspended sediment in the water column; 2) use of Sedflume core data to specify erosion rate parameters in cohesive bed areas; 3) specification of spatially variable bed properties; and 4) inclusion of a sediment bed model that tracks temporal changes in bed composition.

The hydrodynamic model is linked to the sediment transport model via a coupling file, which transfers hydrodynamic transport information (e.g., current velocity, water depth) from the hydrodynamic model to the sediment transport model. For a particular time period, the hydrodynamic model is used to simulate circulation within the Site. During the hydrodynamic simulation, the relevant transport information is output to the coupling file every 15 minutes during the simulation. This frequency of output is necessary to accurately represent the effects of tidal circulation on sediment transport. The coupling file is used as input to the model during a sediment transport simulation. This process significantly reduces the time required to complete a sediment transport simulation because: 1) a larger time-step can be used than if the hydrodynamic and sediment transport models are running in parallel; and 2) the computational burden is lower because the hydrodynamic calculations do not have to be conducted every time a sediment transport simulation is repeated for a specific time period.

The coupling between the hydrodynamic and sediment transport models produces a limitation on the predictive capabilities of the modeling framework. This coupling is one-way, with no feedback between the two models. Changes in bed elevation predicted by the sediment transport model are not incorporated into the hydrodynamic model (i.e., bathymetry in the hydrodynamic model is assumed to remain constant with time). While this limitation may appear to reduce the reliability of the model predications, successful calibration and validation of the model indicate that this limitation in the modeling framework does not have a significant effect on the predictive capabilities of the sediment transport model in the Lower Willamette River.

2.3.2 Sediment Transport Model Theory and Formulation

2.3.2.1 Calculation of Bed Shear Stress

Erosion rate is dependent on bed shear stress, which is calculated using near-bed current velocity predicted by the hydrodynamic model. The bed shear stress calculated within the hydrodynamic model is the total bed shear stress, which represents the total drag on the water column by the sediment bed. The total bed shear stress (τ_{tot}) is the sum of shear stresses associated with skin friction (τ_{sf}) and form drag (τ_{fd}):

$$\tau_{tot} = \tau_{sf} + \tau_{fd} \tag{2-3}$$

Skin friction represents the shear stress generated by sediment particles (i.e., small-scale physical features), whereas form drag corresponds to the drag generated by bedforms (e.g., ripples, dunes) and other large-scale physical features. When simulating erosion, skin friction is considered the dominant component of the bed shear stress for most applications. Thus, it is a reasonable approximation, and a standard approach, to use the skin friction component and neglect form drag for calculating bed shear stress for sediment transport simulations. This approach is consistent with accepted sediment transport theory (Parker 2004). Skin friction shear stress is calculated using the quadratic stress law:

$$\tau_{sf} = \rho_w C_f U^2 \tag{2-4}$$

Where,
ρ_w = the density of water
C_f = the bottom friction coefficient
U = the depth-averaged current velocity.

The bottom friction coefficient is determined using (Parker 2004):

$$C_f = \kappa^2 \ln^{-2}(11 z_{ref} / k_s) \tag{2-5}$$

Where,
z_{ref} = a reference height above the sediment bed
k_s = the effective bed roughness
κ = von Karman's constant (0.4).

The reference height (z_{ref}) is spatially and temporally variable because it is equal to half of the water depth. Thus, the reference height properly incorporates temporal and spatial variations in water depth into the calculation of the bottom friction coefficient. The effective bed roughness is assumed to be proportional to the D_{90} of the surface sediment layer (Parker 2004; Wright and Parker 2004):

$$k_s = 2D_{90} \tag{2-6}$$

Grain size distribution (GSD) data were used to specify D_{90} values for the surface layer of Lower Willamette River sediments. The spatial variability of D_{90} in the Lower Willamette River was evaluated (see Section 2.3.4); accounting for potential spatial variation of D_{90} in the model produces qualitatively correct results (i.e., skin friction increases as bed roughness increases).

The validity of the above approach for calculating the bottom friction coefficient is evaluated as follows. Bottom friction coefficients were calculated for the Lower Willamette River, using representative D_{90} values in the cohesive and non-cohesive bed areas (see Section 2.3.4) over a range water depths (see Table 2-3). The range of bottom friction coefficient values in Table 2-3 is consistent with expected values for cohesive beds (van Rijn 1993). This approach provides an objective method for estimating the effective bed roughness, which will decrease the uncertainty associated with subjective estimates of roughness.

Table 2-3. Bottom Friction Coefficient Values for a Range of Water Depths

Water Depth (m)	Bottom Friction Coefficient: Cohesive Bed (D90 = 280 µm)	Bottom Friction Coefficient: Non-Cohesive Bed (D90 = 1,480 µm)
1	0.0016	0.0024
2	0.0014	0.0020
3	0.0013	0.0018
4	0.0012	0.0017

Notes:
µm = micrometer
m = meter

For use in formulations presented below, a demonstrated accurate equation for bed shear velocity (u*) is defined as (van Rijn 1993):

$$u_* = (\tau_{sf} / \rho_w)^{0.5} \tag{2-7}$$

Current velocity in turbulent flow, which exists in the Lower Willamette River for all flow and tidal conditions, is the sum of two components: time-averaged mean velocity and turbulent fluctuations about the mean value. The bed shear velocity (u*) corresponds to the turbulent-fluctuation component of the current velocity. Thus, the skin friction shear stress is driven by the turbulent fluctuations in the flow, which are randomly variable with time. Random variation in turbulence along the sediment bed is the primary reason that a probabilistic approach to calculating deposition and erosion fluxes is necessary; use of probability of deposition (see Equation 2-8) and suspension (see Equation 2-28) formulations have been incorporated into the model to account for these turbulence effects.

2.3.2.2 Deposition Processes

The deposition flux for size class k sediment (D_k) is expressed as (Ziegler et al. 2000):

$$D_k = P_{dep,k}\, W_{s,k}\, C_k\, \boldsymbol{\Gamma}_k \tag{2-8}$$

where $P_{dep,k}$ is probability of deposition of class k, $W_{s,k}$ is settling speed of class k, C_k is depth-averaged suspended sediment concentration of class k, and $\boldsymbol{\Gamma}_k$ is the stratification correction factor for class k (i.e., ratio between the near-bed suspended sediment concentration and the depth-averaged suspended sediment concentration for class k). Deposition flux has units of mass per unit area per time (e.g., g/cm^2–s). The depth-averaged concentration (C_k) is calculated using the sediment transport model. The ratio $\boldsymbol{\Gamma}_k$ is calculated assuming a known vertical sediment concentration distribution for class k and the current hydrodynamic conditions.

Deposition flux has units of mass per unit area per time (e.g., gram[g]/cm^2–s). The depth-averaged concentration (C_k) is calculated using the sediment transport model.

Probability of deposition of cohesive sediment (i.e., class 1) is determined using the Krone formulation (van Rijn 1993):

$$P_{dep,k} = 1 - (\tau_{sf}/\tau_{cr,dep}) \quad \text{for} \quad \tau_{sf} < \tau_{cr,dep} \qquad (2\text{-}9)$$
$$= 0 \quad \text{for} \quad \tau_{sf} > \tau_{cr,dep}$$

Where
τ_{sf} = bed shear stress (skin friction)
$\tau_{cr,dep}$ = the critical bed shear stress for deposition

The relationship between the probability of deposition and bed shear stress for class 1 is shown in Figure 2-14.

For non-cohesive sediment (i.e., classes 2, 3, and 4), the probability of deposition depends on bed shear stress and particle diameter and is described by a Gaussian distribution (Gessler 1967; Ziegler et al. 2000):

$$P_{dep,k} = (2\pi)^{-0.5} \int EXP(-0.5x^2)\, dx \qquad (2\text{-}10)$$

where the lower and upper limits of the integral are negative infinity and Y, respectively, and EXP corresponds to the exponential function with base e. The parameter Y is given by:

$$Y = 1.75\ (\tau_{c,k}/\tau_{sf} - 1) \qquad (2\text{-}11)$$

Where,
$\tau_{c,k}$ = critical shear stress for suspension of class k sediment, which is:

$$\tau_{c,k} = \rho_w\, u_{*,crs,k}^{\ 2} \qquad (2\text{-}12)$$

Where
$u_{*,crs,k}$ = critical bed shear velocity for initiation of suspension for class k:

$$u_{*,crs,k} = 4\ W_{s,k} / d_{*,k} \qquad \text{for } 1 < d_{*,k} \leq 10 \qquad (2\text{-}13)$$

$$= 0.4\ W_{s,k} \qquad \text{for } d_{*,k} > 10$$

and:

$$d_{*,k} = d_k\ [(s-1)g/\nu^2]^{1/3} \qquad (2\text{-}14)$$

Where,
d_k = particle diameter for class k
s = specific density of particle (i.e., 2.65)
g = acceleration caused by gravity
ν = kinematic viscosity of water.

The non-dimensional particle parameter ($d_{*,k}$) is commonly used in a wide range of sediment transport formulations (van Rijn 1993). The probability of deposition for classes 2 and 3 as a function of bed shear stress and particle diameter is presented in Figure 2-15.

Cohesive (i.e., class 1) particles in the water column aggregate and form flocs. Settling speeds of cohesive flocs have been measured over a large range of concentrations and shear stresses in freshwater (Burban et al. 1990). The Burban settling speed data for cohesive flocs in freshwater were analyzed to develop a formulation to approximate the effects of flocculation on settling speed. This analysis indicated that the settling speed is dependent on the product of the concentration (C_1) and the water column shear stress (G) at which the flocs are formed, resulting in the following relationship:

$$W_{s,1} = 3.3(C_1 G)^{0.12} \qquad (2\text{-}15)$$

where the units of $W_{s,1}$, C_1, and G are meters per day (m/day), milligrams per liter (mg/L), and dynes/cm^2, respectively. For a vertically averaged model, as used in this study, the relevant shear stress for use in Equation 2-15 is the skin friction shear stress (τ_{sf}). The near-bed concentration of suspended cohesive (class 1) particles is represented by the vertically averaged value, which is a valid approximation because of the relatively low settling speed of class 1 particles (i.e., 1 to 10 m/day).

For non-cohesive particles, numerous field and laboratory experiments have demonstrated that a physically realistic representation of the settling speed of a discrete particle is related to the particle diameter, representing size class k, as follows (Cheng 1997):

$$W_{s,k} = (\nu/d_k)\ [(25 + 1.5\ d_{*,k}^{\,2})^{0.5} - 5]^{1.5} \qquad (2\text{-}16)$$

The dependence of settling speed on particle diameter is shown in Figure 2-16.

Applying a 2D, vertically-averaged approximation for the vertical distribution of suspended sediment assumes that particles are uniformly distributed throughout the water column, which is a good approximation for cohesive sediments due to their relatively low settling speeds (i.e., about 1 to 10 meters per day [m/day]). The relatively high settling speeds of suspended sand causes significant vertical stratification to occur, with order of magnitude increases in suspended sediment concentration typically occurring between the top and bottom of the water column. Thus, simulation of suspended sand transport with a vertically averaged model necessitates the use of a correction factor (Γ_k) to incorporate the effects of concentration stratification. Note that Γ for cohesive sediment is always equal to one because of the low settling speed of that type of sediment.

The correction factor relates the vertically averaged concentration of size class k sediment (C_k), which is calculated by the sediment transport model, to the near-bed concentration ($C_{nb,k}$). The vertical distribution of non-cohesive sediment in the water column is calculated using (van Rijn 1984b):

$$C_{V,k}(z) = \begin{cases} C_{nb,k}\left[\left(\frac{a}{h-a}\right)\left(\frac{h}{z}\right) - 1\right] & , \quad \frac{z}{h} < 0.5 \\ C_{nb,k}\left(\frac{a}{h-a}\right)^{\zeta} e^{-4\zeta\left(\frac{z}{h} - 0.5\right)} & , \quad \frac{z}{h} \geq 0.5 \end{cases} \tag{2-17}$$

where:

$C_{V,k}(z)$	=	vertical distribution of suspended sediment concentration, class k*a*
z	=	vertical coordinate ($z = 0$ at sediment bed and $z = h$ at water surface)
a	=	reference height (z=a)
h	=	water depth
ζ	=	suspension parameter defined by (van Rijn 1984b):

$$\zeta = \frac{w_{s,k}}{u_* \, \beta_k} \tag{2-18}$$

where □is□von Karman constant (assumed to be 0.4) and the β-factor, which is related to the vertical diffusion of sediment particles, is given by (van Rijn 1984b):

$$\beta_k 1 + 2\left(\frac{w_{s,k}}{u_*}\right)^2 \quad , \quad 0.1 < \frac{w_{s,k}}{u_*} < 1 \tag{2-19}$$

The vertically-averaged concentration is defined as:

$$C_k = \frac{1}{h}\int_a^h C_{V,k}(z)dz \quad (2\text{-}20)$$

Using Equation 2-17 in the above integral yields:

$$C_k = \frac{C_{nb,k}}{h}\left(\frac{a}{h-a}\right)^{\zeta}\left\{\int_a^{0.5h}\left(\frac{h}{z}-1\right)^{\zeta}dz + \int_{0.5h}^{h} e^{-4\zeta\left(\frac{z}{h}-0.5\right)}dz\right\} \quad (2\text{-}21)$$

The integrals in this equation will be evaluated separately. The first integral does not have a closed form solution. Approximating the solution using the trapezoidal rule and three segments between $z = a$ and $z = 0.5\ h$ (i.e., $\delta z = (0.5h - a)/3$) yields:

$$\int_a^{0.5h}\left(\frac{h}{z}-1\right)^{\zeta}dz = \frac{1}{3}\left[0.5\left(\frac{h}{a}-1\right)^{\zeta} + \left(\frac{h}{a+2\delta z}-1\right)^{\zeta} + 0.5\right] \quad (2\text{-}22)$$

The reference height (a) is calculated using:

$$a = MAX(0.01h, k_{nik}) \quad (2\text{-}23)$$

where h is water depth and k_{nik} is the Nikuradse roughness height:

$$k_{nik} = 33D_{90} \quad (2\text{-}24)$$

The second integral has the following solution:

$$\int_{0.5h}^{h} e^{-4\zeta\left(\frac{z}{h}-0.5\right)}dz = \frac{h}{4\zeta}\left(1-e^{-2\zeta}\right) \quad (2\text{-}25)$$

Inserting Equations 2-22 and 2-25 into Equation 2-21 and solving for $C_{k,nb}$ produces:

$$C_{k,nb} = \Gamma_k C_k \tag{2-26}$$

where:

$$\Gamma_k = \left(\frac{h}{a} - 1\right)^{\zeta} \left\{\frac{h}{4\zeta}\left(1 - e^{-2\zeta}\right) + \frac{1}{3}\left(0.5 - \frac{a}{h}\right) X \left[0.5\left(\frac{h}{a} - 1\right)^{\zeta} + \left(\frac{h}{a+\delta z} - 1\right)^{\zeta} + \left(\frac{h}{a+2\delta z} - 1\right)^{\zeta}\right]\right\} \tag{2-27}$$

2.3.2.3 Erosion Processes: Cohesive Bed

Within sediment bed areas designated as cohesive, the following numerical algorithm is used to calculate the erosion flux of sediment from the bed to the water column, where it is transported as suspended sediment. The erosion flux for size class k sediment (E_k) from a cohesive bed is given by:

$$E_k = \rho_{dry}\, f_{AS,k}\, S_k\, P_{sus,k}\, E_{gross} \tag{2-28}$$

Where,
E_{gross} = the gross erosion rate
$P_{sus,k}$ = probability of suspension for size class k
S_k = the particle-shielding factor for size class k
ρ_{dry} = dry density of bed sediment
$f_{AS,k}$ = the fraction of size class k sediment in the active-surface layer.

Erosion flux has units of mass per unit area per time (e.g., g/cm^2–s).

Erosion of a sediment bed depends on a number of factors, including, but not limited to: shear stress, GSD, dry (bulk) density, total organic carbon (TOC) content, and gas content (Jepsen et al. 1997; Roberts et al. 1998). Factors such as TOC content, gas content, and bioturbation are implicitly incorporated into the cohesive erosion algorithm through the use of site-specific erosion rate data (i.e., Sedflume core data). The rate at which sediment is removed from the consolidated sediment bed and transported to a thin near-bed layer that exists between the consolidated sediment bed and the water column is termed the gross erosion rate (E_{gross}). Some of the eroded sediment in the near-bed layer is re-deposited to the consolidated bed; the rate of re-deposition is referred to as the gross deposition rate (D_{gross}). The remainder of the eroded material in the near-bed layer is transported to the water column; this rate is referred to as the net erosion rate (E_{net}). The near-bed layer discussed above is incorporated into a model of the sediment bed. Erosion rate data obtained from Sedflume testing (Sea Engineering 2006) were analyzed to develop an understanding of the erosion properties of Lower Willamette River cohesive bed sediment (see Section 2.3.4). The goal of that analysis was to develop a functional relationship between E_{gross} and other parameters that affect erosion rate. These relationships and parameters are incorporated into algorithms so that site-specific erosion properties measured for the Lower Willamette River can be represented in the model. Two parameters that affect E_{gross} are shear stress (τ) and bulk density (ρ; Jepsen et al. 1997). An evaluation of Sedflume data indicated minimal correlation exists system-wide between bulk density and erosion rate for Lower Duwamish Waterway sediment. Thus, it is assumed in this study that erosion rate is dependent on skin friction shear stress (Jones and Lick 2001):

$$E_{gross} = A\,\tau_{sf}^{n} \quad \text{for } \tau_{sf} > \tau_{cr} \qquad (2\text{-}29)$$
$$= 0 \quad \text{for } \tau_{sf} \leq \tau_{cr}$$

Where,
E_{gross} = gross erosion rate (cm/s)
τ_{sf} = skin friction shear stress (Pa)
τ_{cr} = critical shear stress (Pa), which is the shear stress at which a small, but measurable, rate of erosion occurs (generally less than 2 millimeters per hour).

The erosion parameters, A and n, are site-specific and may be spatially variable, both horizontally and vertically. Discussion of spatial variations in the erosion parameters in Equation 2-29 is presented in Section 2.3.4.

The erosion rate of each sediment size class is affected by the probability of suspension for that size class ($P_{sus,k}$), which is given by (Jones and Lick 2001):

$$P_{sus,k} = 0 \quad \text{for } \tau_{sf} \leq \tau_{c,k} \qquad (2\text{-}30)$$
$$= [\ln(\beta_1) - \ln(\beta_2)]/[1.39 - \ln(\beta_2)] \quad \text{for } \tau_{sf} \geq \tau \text{ and } \beta_1 \leq 4$$
$$= 1 \quad \text{for } \beta_1 > 4$$

and the non-dimensional parameters are:

$$\beta_1 = u_*/W_{s,k} \qquad (2\text{-}31)$$

$$\beta_2 = u_{*,crs,k}/W_{s,k} \qquad (2\text{-}32)$$

The formulation presented in Equation 2-30 was developed from the results of flume measurements of suspended and bedload transport of sand conducted by Guy et al. (1966). Jones and Lick (2001) analyzed the Guy et al. (1966) data, with Equation 2-30 resulting from their analysis. Probability of suspension as a function of bed shear stress is shown in Figure 2-17 for particle diameters of 130 and 540 µm. This figure shows that for a given shear stress value, the probability of suspension increases with decreasing particle size.

The particle-shielding factor, which is a positive number with a maximum value of one, is used to reduce the erosion flux of smaller particles within a graded bed (i.e., bed with wide range of particle sizes) that are sheltered by larger particles. The particle-shielding factor (S_k) for size class k is formulated as follows (Karim and Kennedy 1981; Rahuel et al. 1989):

$$S_k = (d_k/d_m)^{0.85} \quad \text{for } d_k \leq d_m \tag{2-33}$$

$$= 1 \quad \text{for } d_k > d_m$$

where d_m is the mean particle diameter in the active layer. The relationship between the particle-shielding factor and particle diameter, for three values of mean particle diameter, is shown in Figure 2-18. For a given particle diameter (d_k), the particle-shielding effect increases (i.e., S_k decreases) as the mean particle diameter increases. The particle-shielding factor is consistent with erosion processes within a graded bed, where voids between larger particles provide areas where smaller particles may be shielded (i.e., "hide") from the turbulence at the sediment-water interface that induces erosion. Thus, the particle-shielding factor is a mechanistic parameter that accounts for real processes that affect scour from a graded bed.

The sediment bed model used in the bed scour model is similar to the bed model described in Jones and Lick (2001). This bed model has been developed over the previous 20 years and used within the SEDZL and SEDZLJ algorithms (Ziegler and Lick 1988; Ziegler et al. 2000; Jones and Lick 2001; QEA 2008). The SEDZL/SEDZLJ bed model has been successfully used in over 30 sediment transport modeling studies, including: Lower Duwamish Waterway (Washington), Upper Hudson River (New York), Lavaca Bay (Texas), Grasse River (New York), Upper Mississippi River (Minnesota), Watts Bar Reservoir/Tennessee River (Tennessee), and Patrick Bayou (Texas).

A multi-layer bed model is used in the SEDZLJ algorithm, with each bed layer having specific erosion rate parameters (i.e., τ_{cr}, A, and n). For this study, five bed layers are used, with the initial thickness of each layer being 5 cm. Use of 5-cm layers in the bed model is based on the vertical variation in erosion rate data obtained from the Sedflume cores; the shear stress series were repeated in approximately 5-cm increments in a core during the Sedflume tests. Discretizing the bed into five layers allows specifying vertical variation in erosion properties, with the erodibility of cohesive sediment generally decreasing with depth in the bed, primarily due to consolidation processes. Additional discussion about the 5-layer bed model is presented in Section 2.3.4.

The effects of consolidation on erosion properties of deposited sediment are not explicitly incorporated into the bed model. If the initial layer 1 is present, then deposited sediment is added to layer 1 (i.e., surface layer) of the bed model and, thus, that sediment has the same erosion properties as the surface layer. If the initial layer 1 is not present (i.e., that layer has been eroded), then a new surface layer is created by the deposited sediment, which has the same erosion properties as the initial layer 1. This approach produces conservative results during a high-flow event because the erosion properties of sediment deposited prior to the event will not have been reduced due to consolidation.

Erosion from cohesive and non-cohesive beds is affected by bed armoring, which is a process that tends to limit the amount of bed scour during a high-flow event. Bed armoring occurs in a bed that contains a range of particle sizes (e.g., clay, silt, sand). During a high-flow event when erosion is occurring, finer particles (i.e., clay and silt) tend to be eroded at a faster rate than coarser particles (i.e., sand). The differences in erosion rates of various particle sizes creates a thin layer at the surface of the bed, referred to as the active layer, that is depleted of finer particles and enriched with coarser particles. This depletion-enrichment process can lead to bed armoring, where the active layer is primarily composed of coarse particles that have limited mobility.

After bed armoring occurs during a high-flow event, various physical mixing processes in the surface layer of the bed (e.g., bioturbation, ship-induced resuspension) can affect the armor layer. The effects of physical mixing processes on bed armoring are not well understood at the present time; these effects are not explicitly incorporated into the bed model and bed armoring algorithm. However, the effects of physical mixing processes are implicitly included into the bed model through use of the Sedflume data, which incorporates these effects into the erosion rate data. Physical mixing in the surface layer is one reason why near-surface sediment is generally more erodible than deeper sediment.

The bed armoring process is simulated using an active layer at the surface of the bed, with the gross erosion rate being affected by the composition of the active layer (Jones and Lick 2001). The active layer is a theoretical construct that approximates the near-bed layer mentioned during the description of gross deposition and erosion rates previously in this section. The active layer is part of a numerical algorithm and it was created as a "holding area," such that the bed model realistically represents the complex processes at the sediment-water interface. Even though the active-layer approach used in the model is a simplification of various complex processes, it is conceptually realistic and has been shown to produce accurate results in previous modeling studies.

In this study, four size classes of sediment were used. Class 1 sediment represents cohesive sediment (i.e., clay and silt, less than 62 μm diameter). Class 2 sediment represents fine sand (i.e., 62 to 250 μm diameter). Class 3 sediment represents medium and coarse sand (i.e., 250 to 2,000 μm diameter). Class 4 sediment represents gravel (i.e., greater than 2,000 μm diameter). The bed model tracks changes in the composition of the active layer associated with erosion and deposition; temporal changes in active layer composition affect the erosion process.

The surface layer in the bed model (i.e., top 5-cm layer) is divided into two zones: 1) active layer; and 2) parent bed. The active layer is at the top of the surface layer and the parent bed is below it. The active layer interacts with the water column; erosion and deposition across the sediment-water interface occurs in the active layer. Use of an active layer to simulate the effects of bed armoring is frequently used in sediment transport models (Rahuel et al. 1989). The active layer is composed of two sub-layers (QEA 2008): 1) active-surface layer; and 2) active-buffer layer. The active-surface layer interacts with the water column, while the active-buffer layer controls interactions between the active-surface layer and the parent bed (Figures 2-19 and 2-20).

The thickness of the active-surface layer is assumed to depend on bed shear stress and GSD. The formulation used to calculate active-surface layer thickness (T_{AS}) is (Jones and Lick 2001):

$$T_{AS} = 2\ d_m\ (\tau_{sf} / \tau_{cr}) \tag{2-34}$$

Where,
d_m = the mean particle diameter in the active layer.

The active-surface layer thickness is temporally and spatially variable, and it changes as the composition of the bed and bed shear stress change with time. The active-surface layer thickness is determined using Equation 2-34, with the bed model tracking the mass per unit area using:

$$M_{AS} = \rho_{dry}\ T_{AS} \tag{2-35}$$

Where,
M_{AS} = the total sediment mass per unit area in the active-surface layer
ρ_{dry} = the dry density of bed sediment.

The thickness, or mass per unit area, of the active-surface layer changes with time as T_{AS} changes as a result of increases or decreases in mean particle diameter or bed shear stress. Let δ_{SB} represent changes in active-surface layer mass, for size class k, caused by temporal changes in M_{AS}. Expansion and contraction of the active-surface thickness (i.e., T_{AS}) causes interactions between the active-surface and active-buffer layers, which result in mass transfer between the two layers. For increasing M_{AS} (i.e., $M_{AS}^{N+1} > M_{AS}^{N}$, where the superscript N represent time-level N in the numerical model):

$$\delta_{SB,k} = f_{AB,k}\ (M_{AS}^{N+1} - M_{AS}^{N}) \tag{2-36}$$

Where,
$f_{AB,k}$ = the fraction of size class k sediment in the active-buffer layer.

For decreasing or constant M_{AS} (i.e., $M_{AS}^{N+1} \le M_{AS}^{N}$):

$$\delta_{SB,k} = f_{AS,k}\ (M_{AS}^{N+1} - M_{AS}^{N}) \tag{2-37}$$

Where
$f_{AS,k}$ = the fraction of size class k sediment in the active-surface layer.

The change in active-surface layer mass is calculated using:

$$M_{AS,k}^{N+1} = M_{AS,k}^{N} + \delta_{SB,k} + \Delta t\,(D_k - E_k - f_{AS,k}\,D_{tot} + f_{AB,k}\,E_{tot}) \qquad (2\text{-}38)$$

Where,
$M_{AS,k}$ = active-surface layer mass per unit area for size class k sediment
E_k = the erosion flux for size class k sediment
D_k = the deposition flux for size class k sediment
Δt = the numerical time-step.

The total deposition and erosion fluxes are given by:

$$D_{tot} = \Sigma\, D_k \qquad (2\text{-}39)$$

$$E_{tot} = \Sigma\, E_k \qquad (2\text{-}40)$$

where the summations are over the four size classes. In Equation 2-40, the values of E_k are calculated using Equation 2-28 for each size class k. Thus, E_{tot} is affected by the composition of the active-surface layer. Note that the deposition and erosion flux terms in the parentheses on the right-hand side of Equation 2-38 do not sum to zero for a specific size class k. This characteristic of the algorithm generates bed armoring effects due to unequal mass transfer of different sediment size classes between the active-surface, active-buffer, and parent-bed layers. However, conservation of mass is assured when Equation 2-38 is summed over all sediment size classes, which results in the sum of the deposition and erosion flux terms being equal to zero.

The terms on the right-hand-side of Equation 2-38 correspond to the following changes in the mass of the active-surface layer: 1) $\delta_{SB,k}$ is an increase in mass of class k sediment if the total active-surface layer mass is increasing (i.e., mass added from active-buffer layer) and it is a decrease in mass of class k sediment if the total active-surface layer mass is decreasing (i.e., mass lost to active-buffer layer); 2) $\Delta t\, D_k$ is an increase in mass of class k sediment due to deposition from the water column to the bed; 3) $\Delta t\, E_k$ is a decrease in mass of class k sediment due to erosion from the bed to the water column; 4) $\Delta t\, f_{AS,k}\, D_{tot}$ is a decrease in mass of class k sediment caused by movement of sediment from the active-surface layer to the active-buffer layer due to deposition; and 5) $\Delta t\, f_{AB,k}\, E_{tot}$ is an increase in mass of class k sediment caused by movement of sediment from the active-buffer layer to the active-surface layer due to erosion (see Figure 2-19).

The change in active-buffer layer mass for size class k ($M_{AB,k}$) is calculated using:

$$M_{AB,k}^{N+1} = M_{AB,k}^{N} - \delta_{SB,k} + \Delta t\,[(f_{AS,k} - f_{AB,k})D_{tot} - f_{AB,k}\,E_{tot}] \tag{2-41}$$

It is assumed that there is no mass transfer between the buffer layer and the parent bed due to erosion processes. The terms on the right-hand-side of Equation 2-41 correspond to the following changes in the mass of the active-buffer layer: 1) $\delta_{SB,k}$ is a decrease in mass of class k sediment if the total active-surface layer mass is increasing (i.e., mass lost to active-surface layer), and it is an increase in mass of class k sediment if the total active-surface layer mass is decreasing (i.e., mass added from active-surface layer); 2) $\Delta t\ f_{AS,k}\ D_{tot}$ is an increase in mass of class k sediment caused by movement of sediment from the active-surface layer to the active-buffer layer due to deposition; 3) $\Delta t\ f_{AB,k}\ D_{tot}$ is a decrease in mass of class k sediment caused by movement of sediment from the active-buffer layer to the parent-bed layer due to deposition; and 4) $\Delta t\ f_{AB,k}\ E_{tot}$ is a decrease in mass of class k sediment caused by movement of sediment from the active-buffer layer to the active-surface layer due to erosion.

When the buffer layer is depleted of sediment (typically during an erosion event), the active-surface layer interacts directly with the parent bed (Figure 2-20). Let $\delta_{SP,k}$ represent changes in active-surface layer mass, for size class k, caused by temporal changes in M_{AS} and expansion/contraction interactions between the active-surface and parent-bed layers. For increasing M_{AS}:

$$\delta_{SP,k} = f_{P,k}\,(M_{AS}^{N+1} - M_{AS}^{N}) \tag{2-42}$$

where $f_{P,k}$ is the fraction of size class k sediment in the parent-bed layer. For decreasing or constant M_{AS}:

$$\delta_{SP,k} = f_{AS,k}\,(M_{AS}^{N+1} - M_{AS}^{N}) \tag{2-43}$$

The change in active-surface layer mass for size class k is calculated using:

$$M_{AS,k}^{N+1} = M_{AS,k}^{N} + \delta_{SP,k} + \Delta t\,(D_k - E_k - f_{As,k}\,D_{tot} + f_{P,k}\,E_{tot}) \tag{2-44}$$

The terms on the right-hand-side of Equation 2-44 correspond to the following changes in the mass of the active-surface layer: 1) $\delta_{SP,k}$ is an increase in mass of class k sediment if the total active-surface layer mass is increasing (i.e., mass added from parent-bed layer) and it is a decrease in mass of class k sediment if the total active-surface layer mass is decreasing (i.e., mass lost to parent-bed layer); 2) $\Delta t\ D_k$ is an increase in mass of class k sediment due to deposition from the water column to the bed; 3) $\Delta t\ E_k$ is an decrease in mass of class k sediment due to erosion from the bed to the water column; 4) $\Delta t\ f_{AS,k}\ D_{tot}$ is a decrease in mass of class k sediment caused by movement of sediment from the active-surface layer to the parent-bed layer due to deposition; and 5) $\Delta t\ f_{P,k}\ E_{tot}$ is an increase in mass of class k sediment caused by movement of sediment from the parent-bed layer to the active-surface layer due to erosion.

The change in parent-bed layer mass for size class k ($M_{P,k}$) is determined from:

$$M_{P,k}^{N+1} = M_{P,k}^{N} - \delta_{SP,k} + \Delta t \left(f_{AS,k} D_{tot} - f_{P,k}\, E_{tot}\right) \tag{2-45}$$

The terms on the right-hand-side of Equation 2-45 correspond to the following changes in the mass of the parent-bed layer: 1) $\delta_{SP,k}$ is a decrease in mass of class k sediment if the total active-surface layer mass is increasing (i.e., mass lost to active-surface layer) and it is an increase in mass of class k sediment if the total active-surface layer mass is decreasing (i.e., mass added from active-surface layer); 2) $\Delta t\ f_{AS,k}\ D_{tot}$ is an increase in mass of class k sediment caused by movement of sediment from the active-surface layer to the parent-bed layer due to deposition; and 3) $\Delta t\ f_{P,k}\ E_{tot}$ is a decrease in mass of class k sediment caused by movement of sediment from the parent-bed layer to the active-surface layer due to erosion.

After the buffer layer is depleted, a new active-buffer layer is created when the active-surface layer decreases in thickness as a result of decreasing bed shear stress. For the condition when $M_{AB,k}^{N+1}$ equals zero and M_{AS} is decreasing (i.e., $M_{AS}^{N+1} \leq M_{AS}^{N}$), then the initial mass of the new active-buffer layer, for size class k, is:

$$M_{AB,k}^{N+1} = f_{P,k} \left(M_{AS}^{N} - M_{AS}^{N+1}\right) \tag{2-46}$$

This amount of mass is removed from the parent-bed layer, so that mass is conserved.

The fractions of each sediment size class are updated after the new sediment masses are calculated in each layer:

$$f_{AS,k} = M_{AS,k}^{N+1} / M_{AS}^{N+1} \tag{2-47}$$

$$f_{AB,k} = M_{AB,k}^{N+1} / M_{AB}^{N+1} \tag{2-48}$$

$$f_{P,k} = M_{P,k}^{N+1} / M_{P}^{N+1} \tag{2-49}$$

where M_{AS}^{N+1}, M_{AB}^{N+1},and M_{P}^{N+1} are total sediment mass per unit area in the active-surface, active-buffer, and parent-bed layers, respectively.

The numerical algorithm presented above for the interactions between the active-surface, active-buffer, and parent-bed layers may be difficult to understand from a conceptual viewpoint. The following sequence of figures is intended to clarify the mechanistic interactions between the three layers due to temporal variations in bed shear stress, which result in expansion and contraction of the active layer. It is assumed that initially (i.e., time = t_1) two layers exist: 1) active-surface layer (with thickness $T_{AS,1}$ corresponding to a shear stress value of $\tau_{sf,1}$); and 2) parent-bed layer (see Figure 2-21). As the shear stress increases to $\tau_{sf,2}$ (which is greater than $\tau_{sf,1}$) at time = t_2, the active-surface layer thickness increases to $T_{AS,2}$, and sediment is transferred from the parent-bed layer to the active-surface layer (Figure 2-22). The shear stress reaches a maximum value at time = t_2 and decreases to a value of $\tau_{sf,3}$ at time = t_3. As the shear stress decreases during this time interval (i.e., t_2 to t_3), an active-buffer layer is created as the active-surface layer contracts in size, which is the process that generates an active-buffer layer (Figure 2-23). This new active-buffer layer was created from a portion of the active-surface layer that existed at time = t_2; sediment was transferred from the active-surface layer to the active-buffer layer. As the shear stress continues to decrease during the time interval between t_3 and t_4, the active-surface and active-buffer layers decrease and increase in thickness, respectively (Figure 2-24). The shear increases during the time interval between t_4 and t_5, which causes sediment to be transferred from the active-buffer layer (which is contracting) to the active-surface layer (which is expanding; see Figure 2-25). Note that during the time interval between t_2 and t_5, when the shear stress is less than the maximum value of $\tau_{sf,2}$, the sum of the thicknesses of the active-surface and active-buffer layers remains constant at a value of $T_{AS,2}$ (assuming that no deposition or erosion occurs). During the time interval between t_5 and t_6, the active-buffer layer is destroyed, and sediment is transferred from the parent-bed layer to the active-surface layer, as the shear stress exceeds the original maximum value of $\tau_{sf,2}$ and the active-surface layer expands to a thickness greater than $T_{AS,2}$ (Figure 2-26). As the shear stress decreases from the new maximum value of $\tau_{sf,6}$, a new active-buffer layer is created from the active-surface layer as that layer contracts in size (Figure 2-27).

The structure of the bed model described above is based on heuristic concepts that were developed from a general understanding of cohesive bed processes. The overall concepts applied to the model and the general behavior of the model are consistent with known processes. However, uncertainty exists in some details of the model structure (e.g., transfer of sediment between the active-surface, active-buffer, and parent-bed layers as the active layer expands and contracts). Due to the complexity of the model structure, a unique methodology does not exist, and a wide range of alternatives can be constructed from proposed general structure. However, the approach that is described above, and used in the Lower Willamette River modeling study, is consistent with a general understanding of cohesive bed processes, and it does produce reasonable results.

2.3.2.4 Erosion Processes: Non-Cohesive Bed

Non-cohesive sediment bed transport is dominated by gravitational, lift, and drag forces acting on individual particles. Cohesive forces are negligible compared to these other forces and are not evident in non-cohesive bed behavior. Non-cohesive beds generally contain only a small amount of clay and silt particles. Numerous laboratory and field studies have been conducted on the erosion properties of non-cohesive sediments; see van Rijn (1993) for an overview. These investigations have led to the development of various formulations for quantification of non-cohesive suspended and bedload transport. Several investigators have evaluated the accuracy of different quantitative approaches using laboratory and field data (Garcia and Parker 1991; Voogt et al. 1991; van den Berg and van Gelder 1993). The results of these investigations have shown that the formulations developed by van Rijn (1984a, b, c) provide one of the best methods for calculating suspended load transport of non-cohesive sediments. The van Rijn equations have been successfully used in sediment transport modeling studies of riverine (Ziegler et al. 2000) and estuarine (van Rijn et al. 1990) systems over a wide range of flow and sediment conditions.

The numerical algorithm discussed below is used to calculate the erosion flux of sediment from a non-cohesive bed to the water column, where it is transported as suspended sediment. Following the van Rijn method, the equations presented below are used to calculate the erosion flux for sediment size class k, which is represented by an effective particle diameter (d_k). The critical bed shear velocity for initiation of bedload transport ($u_{*,crb,k}$) is calculated using the Shields criteria (see Figure 2-28):

$$u_{*,crb,k} = [(s-1)\ g\ d_k\ \theta_{cr,k}]^{0.5} \tag{2-50}$$

where θ_{cr} is the critical mobility parameter, which is approximated by (van Rijn 1993):

$$\begin{aligned} \theta_{cr,k} &= 0.24\ d_{*,k}^{-1} && \text{for } d_{*,k} \leq 4 \\ &= 0.14\ d_{*,k}^{-0.64} && \text{for } 4 < d_{*,k} < 10 \\ &= 0.04\ d_{*,k}^{-0.10} && \text{for } 10 < d_{*,k} < 20 \\ &= 0.013\ d_{*,k}^{0.29} && \text{for } 20 < d_{*,k} < 150 \\ &= 0.055 && \text{for } d_{*,k} > 160 \end{aligned} \tag{2-51}$$

and $d_{*,k}$ is calculated using Equation 2-14. Equation 2-51 is a piece-wise fit to the Shields curve that was developed by van Rijn (1993). Critical shear stresses for initiation of bedload ($\tau_{crb,k}$) and suspended load ($\tau_{crs,k}$) transport are calculated as follows:

$$\tau_{crb,k} = \rho_w\ u_{*,crb,k}^2 \tag{2-52}$$

$$\tau_{crs,k} = \rho_w\ u_{*,crs,k}^2 \tag{2-53}$$

The relationships between particle diameter and the critical bed shear stresses for bedload and suspended load transport are shown in Figure 2-29. For sediment class 1, which represents clay and silt, it is assumed that Equations 2-13 and 2-52 through 2-53 can be extrapolated to particle sizes less than 62 µm (i.e., d* less than 1.47). This assumption is commonly used for simulation of non-cohesive sediment transport with a graded bed (i.e., mixture of sediment particle sizes), and it has a minimal effect on model predictions in non-cohesive bed areas.

If the bed shear stress exceeds the critical shear stress for suspended load transport, then the equilibrium sediment concentration ($C_{eq,k}$) at a reference height (z = a) above the bed is calculated using:

$$C_{eq,k} = 0.015\ (d_k\ T_k^{\ 1.5}) / (a\ d_{*,k}^{\ 0.3}) \tag{2-54}$$

where T_k is the transport stage parameter, given by:

$$T_k = (u_*/u_{*,crs,k})^2 - 1 \qquad \text{for} \qquad u_* > u_{*,crs,k} \tag{2-55}$$

The reference height (a) is calculated using:

$$a = \mathrm{MAX}\ (0.01\ h,\ k_{nik}) \tag{2-56}$$

where h is water depth and k_{nik} is the Nikuradse roughness height:

$$k_{nik} = 33\ D_{90} \tag{2-57}$$

The erosion flux for size class k sediment for a non-armoring sediment bed is calculated using:

$$E_{na,k} = -\ W_{s,k}\ (C_{a,k} - Ceq,k) \qquad \text{for} \qquad C_{a,k} < C_{eq,k} \tag{2-58}$$

where $C_{a,k}$ is the suspended sediment concentration of size class k at z = a. For the 3D model, $C_{a,k}$ is set equal to the suspended sediment concentration, as predicted by the water-column transport model, in the first grid cell above the bed.

Similar to the cohesive bed discussed in Section 2.3.2.3, bed armoring processes occur in the non-cohesive bed and those processes affect the erosion flux from that bed type. An active layer is assumed to exist at the surface of the non-cohesive bed, with the thickness of that layer calculated using the following relationship:

$$T_{AS,non} = 2\ d_{50}\ (\tau_{sf} / \tau_{crb})^N \tag{2-59}$$

Where,
T = active layer thickness in the non-cohesive bed
d_{50} = median particle diameter
τ_{crb} = critical shear stress for initiation of bedload transport
N = an exponent that ranges between 0.1 and 1

A bed model tracks changes in the composition of the non-cohesive active layer associated with erosion and deposition, as well as interactions between the active and parent bed layers. Thus, the erosion flux for size class k sediment from an armoring bed ($E_{non,k}$) is given by:

$$E_{non,k} = f_{non,a,k} S_k E_{na,k} \qquad (2\text{-}60)$$

Where,
$f_{non,a,k}$ = the fraction of class k sediment in the active layer of the non-cohesive bed
S_k = the particle-shielding factor (see Equation 2-33).

The particle-shielding factor (Karim and Kennedy 1981; Rahuel et al. 1989) was included in the erosion flux for an armoring bed because this factor accounts for the effects of differential erosion rates.

2.3.3 Sediment Transport Model Inputs: Sediment Properties

Inputs for the sediment transport model are separated into three broad categories: 1) sediment properties (Section 2.3.3); 2) bed properties (Section 2.3.4); and 3) boundary conditions (Section 2.3.5). Sediment properties correspond to the physical properties of sediment particles (i.e., effective particle diameter, settling speed). Bed properties range from bulk bed characteristics (e.g., dry density, GSD) to erosion rates. Determining boundary conditions for the model corresponds to the specification of sediment loads at different inflow locations.

For freshwater tidal rivers such as the Lower Willamette River, suspended sediment particles typically have a range of sizes, with particle diameters ranging from less than 1 μm clays to coarse sands on the order of 1,000 μm (van Rijn 1993). Simulation of the entire particle size spectrum is impractical for several reasons: simulation times and array-storage requirements increase with each particle size class that is added, limitations in GSD data for the sediment bed make it difficult to specify initial conditions for the entire spectrum, and sparse data for the composition of the external sediment load make it problematic for specifying this boundary condition for the entire spectrum. Therefore, particles were separated into four classes: 1) clay and silt with particle diameters less than 62 μm; 2) fine sand (62 to 250 μm); 3) medium and coarse sand (250 to 2,000 μm); and gravel (greater than 2,000 μm). Use of these four size classes provides an adequate approximation of the GSD of bed sediment observed in the Lower Willamette River for achieving the objectives of this study. Based on experience from previous modeling studies (e.g., QEA 2008), the four size classes used in Lower Willamette River simulations provide a realistic range of sediment particle sizes (from clay to gravel) that are present in the graded bed of the river. Inclusion of this wide range of particle sizes in the model is necessary for simulation of bed armoring processes during an erosion event. From a practical point of view, simulating the transport of four sediment size classes makes it possible to conduct long-term, multi-year simulations in a practical amount of time. Finally, the results of the model calibration and validation exercises (discussed below) indicate that use of four sediment size classes is sufficient for producing a modeling framework with adequate accuracy and reliability for the application and use of the Lower Willamette River sediment transport model.

For convenience, the four sediment classes have been labeled as noted in Table 2-4. Each sediment size class is represented as an effective particle diameter. The effective particle diameter for classes 1, 2, and 3 were treated as adjustable calibration parameters, see Section 2.3.6. Prior to model calibration, effective particle diameters for classes 2 through 4 were estimated using the following approach (QEA 2008). GSD data are available for the surface layer of sediment cores collected from the Lower Willamette River. For sand and gravel particles, the GSD data were reported as the fractional composition for these seven particle size ranges: 1) 72 to 106 µm; 2) 106 to 250 µm; 3) 250 to 425 µm; 4) 425 to 850 µm; 5) 850 to 2,000 µm; 6) 2,000 to 4,750 µm; and 7) greater than 4,750 µm. The effective diameter of each size range corresponds to the geometric mean of that range. For example, the geometric mean of the 106 to 250 µm size range is 163 µm. The GSD data provide information on the relative amounts of sand in each of the seven size ranges. The effective diameters of classes 2, 3, and 4 (i.e., d2, d3, d4) for a particular core sample were estimated from the GSD data using (QEA 2008):

$$d_2 = (f_{r1}G_{r1} + f_{r2}G_{r2}) / (f_{r1} + f_{r2}) \quad (2\text{-}61)$$

$$d_3 = (f_{r3}G_{r3} + f_{r4}G_{r4} + f_{r5}G_{r5}) / (f_{r3} + f_{r4} + f_{r5}) \quad (2\text{-}62)$$

$$d_4 = (f_{r6}G_{r6} + f_{r7}G_{r7}) / (f_{r6} + f_{r7}) \quad (2\text{-}63)$$

Where,
f_{rk} = fractional content of size range k
G_{rk} = geometric mean diameter of size range k.

Using Equations 2-61 through 2-63 to analyze the GSD data for surface-layer sediment produced median values of effective diameters for classes 2, 3, and 4 as listed in Table 2-4.

Table 2-4. Characteristics of Sediment Particle Size Classes

Sediment Size Class	Particle Size Range (µm)	Data-Based Effective Particle Diameter (µm)	Calibration Effective Particle Diameter (µm)
1: clay and silt	< 62	Not Applicable	Not Applicable
2: fine sand	62 – 250	140	100
3: medium, coarse sand	250 – 2,000	550	700
4: gravel	> 2,000	2,800	2,800

Notes:
µm = micrometer

2.3.4 Sediment Transport Model Inputs: Bed Properties

The sediment bed in the Lower Willamette River was separated into three distinct types: 1) cohesive (i.e., muddy bed composed of a mixture of clay, silt, sand, and organic matter); 2) non-cohesive (i.e., sandy bed composed of sand and gravel, with small amounts of clay and silt); and 3) hard bottom (i.e., no erosion or deposition). Delineation of the sediment bed into cohesive, non-cohesive and hard bottom areas was accomplished using GSD data from sediment cores collected during the GeoSea and Round 2 field studies during 2000 and 2004, respectively (GeoSea 2001; Integral 2005a, 2005b, 2006). GSD data were available at a total of 1,187 locations in the Study Area (see Figures 2-30 and 2-31). Sediment cores were classified as cohesive using the following criteria: 1) median particle diameter (D_{50}) less than 250 µm; and 2) clay/silt content greater than 15 percent (Ziegler and Nisbet 1994). It was assumed that the sediment bed was hard bottom in the following areas: 1) upstream of RM 12.9 in the Lower Willamette River; 2) Multnomah Channel; and 3) Columbia River. The bed map for the Study Area is shown on Figure 2-32. About 81 percent of the bed area between RM 2 and 11 is cohesive.

The sediment transport model requires specification of the following bed property inputs within the Lower Willamette River: 1) dry (bulk) density; 2) initial sediment bed composition (i.e., relative amounts of sediment size classes 1, 2, 3, and 4); 3) median particle diameter (D_{50}); 4) effective bed roughness (which is proportional to D_{90}, see Equation 2-6); and 5) erosion rate properties in cohesive bed areas. The dry density of the bed was assumed to be spatially variable within the Lower Willamette River, with different values in the cohesive and non-cohesive bed areas. For cohesive bed areas, the dry density has a value of 0.72 g/cm^3, which corresponds to the average value of 596 samples. For non-cohesive bed areas, the dry density has a value of 1.2 g/cm^3, which corresponds to the average value of 162 samples. Dry density is assumed to be horizontally and vertically constant within all areas of a particular bed type.

Spatial distributions of D_{50} and D_{90} values were developed from the GSD data collected at 1,187 locations in the Study Area, see Figure 2-33. The proportional content of the four sediment size classes (i.e., classes 1, 2, 3 and 4) in the bed must be specified at each grid cell at the beginning of a simulation. Spatial distributions of bed composition were specified as initial conditions for the sediment transport model using the GSD data, see Figures 2-34 and 2-35. Changes in bed composition due to erosion and deposition processes are predicted by the model at each grid cell during a simulation. As a reference Table 2-5 presents the average values of D_{50}, D_{90}, and composition of the bed for cohesive and non-cohesive areas.

Table 2-5. Average Values for Bed Properties Initial Conditions

Bed Type	D_{50} (µm)	D_{90} (µm)	Class 1 Content (%)	Class 2 Content (%)	Class 3 Content (%)	Class 4 Content (%)
Cohesive	50	280	64	26	9	1
Non-Cohesive	510	1,480	13	14	64	9

Notes:
µm = micrometer

A Sedflume study was conducted during 2006 to obtain data on the erosion properties of Lower Willamette River sediments. Cores were collected from 19 locations (Figure 2-36). Details of the field study, including core collection and processing, are described in Sea Engineering (2006). Erosion rates as a function of depth in the bed and applied shear stress were measured over the top 30 cm of each core using Sedflume. Sediment samples were also obtained at 5-cm intervals from each core and analyzed for bulk (wet) density and GSD.

Erosion rate data obtained from Sedflume testing were analyzed to develop an understanding of the erosion properties of Lower Willamette River sediments in cohesive bed areas. The goal of this analysis was to develop a functional relationship between Egross and bed shear stress. The site-specific parameters in Equation 2-29 (i.e., A, n, τ_{cr}) were determined using the erosion rate data collected during the Sedflume field study. Four of the 19 Sedflume cores (i.e., cores SF-2, SF-6, SF-7, SF-18) were determined to consist of non-cohesive (i.e., sandy) sediment and those cores were not included in the analysis described below; Sedflume erosion rate data are only applicable to cohesive bed sediment.

The erosion rate properties of the 15 cores were analyzed using the following procedure. Each core was divided into five layers, with these layers representing the following depth intervals: 0 to 5, 5 to 10, 10 to 15, 15 to 20, and 20 to 25 cm. These depth intervals were chosen because the shear stress series used in the Sedflume tests, where shear stress was increased from low to high values, were cycled over approximately 5-cm thick layers. The erosion rate data within each layer of a particular core were analyzed through application of a log-linear regression analysis between erosion rate and shear stress. The log-linear regression analysis produced values of A and n for each layer in a particular core. The results of this analysis for the Sedflume cores with cohesive sediment are presented in Figures 2-37 through 2-51. The critical shear stress for each 5-cm layer was calculated using:

$$\tau_{cr} = (E_{cr} / A)^{1/n} \tag{2-64}$$

Where,
E_{cr} = 0.0001 cm/s.

The erosion rate parameters (i.e., A, n, τ_{cr}) for each core within the five depth intervals are listed in Tables 2-6 through 2-10. Note that the values of A and n in these tables correspond to units of cm/s for E_{gross} and pascal (Pa) for bed shear stress in Equation 2-29. The correlation coefficient (R^2) values presented in the tables are from the log-linear regression analysis, with perfect correlation corresponding to an R^2 value of one.

Table 2-6. Erosion Rate Parameters for 0 to 5 cm Layer

Sediment Core ID	River Mile Location	Proportionality Constant: A	Exponent: n	Correlation Coefficient (R^2)	Critical Shear Stress (Pa)
SF-1	2.4	0.00113	2.4	0.97	0.36
SF-3	3.7	0.00504	1.6	0.96	0.09
SF-4	4.0	0.00244	2.3	0.99	0.25
SF-5	4.8	0.00137	2.0	0.94	0.27
SF-8	6.1	0.00473	2.7	0.98	0.24
SF-9	6.4	0.00081	2.0	0.80	0.35
SF-10	6.8	0.00110	2.25	0.95	0.33
SF-11	6.9	0.00025	3.1	0.98	0.73
SF-12	7.6	0.00430	1.6	0.92	0.10
SF-13	8.0	0.00218	1.3	0.76	0.10
SF-14	8.3	0.00140	1.3	0.76	0.14
SF-15	8.6	0.00546	2.1	0.96	0.15
SF-16	9.3	0.00065	2.6	0.90	0.49
SF-17	10.0	0.00061	2.9	0.96	0.54
SF-19	10.4	0.00115	2.3	0.95	0.34

Table 2-7. Erosion Rate Parameters for 5 to 10 cm Layer

Sediment Core ID	River Mile Location	Proportionality Constant: A	Exponent: n	Correlation Coefficient (R^2)	Critical Shear Stress (Pa)
SF-1	2.4	0.00106	2.8	0.99	0.43
SF-3	3.7	0.00056	4.6	0.99	0.69
SF-4	4.0	0.00043	3.7	0.97	0.67
SF-5	4.8	0.00014	3.2	0.96	0.89
SF-8	6.1	0.00151	2.1	0.96	0.28
SF-9	6.4	0.00015	3.1	0.99	0.86
SF-10	6.8	0.00036	3.1	0.99	0.66
SF-11	6.9	0.00002	4.4	0.97	1.33
SF-12	7.6	0.00054	2.4	0.99	0.49
SF-13	8.0	0.00115	2.6	0.95	0.38
SF-14	8.3	0.00014	1.9	0.83	0.83
SF-15	8.6	0.00117	2.1	0.96	0.32
SF-16	9.3	0.00306	2.0	0.93	0.18
SF-17	10.0	0.00047	3.0	0.98	0.59
SF-19	10.4	0.00120	2.2	0.60	0.32

Table 2-8. Erosion Rate Parameters for 10 to 15 cm Layer

Sediment Core ID	River Mile Location	Proportionality Constant: A	Exponent: n	Correlation Coefficient (R^2)	Critical Shear Stress (Pa)
SF-1	2.4	0.00048	3.9	0.98	0.67
SF-3	3.7	0.00608	2.8	0.98	0.23
SF-4	4.0	0.00034	2.8	0.99	0.64
SF-5	4.8	0.00026	2.6	0.99	0.68
SF-8	6.1	N/A	N/A	N/A	N/A
SF-9	6.4	0.00039	2.3	0.93	0.56
SF-10	6.8	0.00008	3.0	0.95	1.08
SF-11	6.9	0.00358	1.7	0.89	0.12
SF-12	7.6	0.00132	1.8	0.99	0.23
SF-13	8.0	0.00030	2.7	0.90	0.66
SF-14	8.3	0.00003	2.8	0.94	1.47
SF-15	8.6	0.00039	3.3	0.97	0.66
SF-16	9.3	0.00163	2.8	0.94	0.37
SF-17	10.0	0.00040	3.0	0.93	0.63
SF-19	10.4	0.00088	2.9	0.84	0.47

Table 2-9. Erosion Rate Parameters for 15 to 20 cm Layer

Sediment Core ID	River Mile Location	Proportionality Constant: A	Exponent: n	Correlation Coefficient (R^2)	Critical Shear Stress (Pa)
SF-1	2.4	0.00097	2.4	0.99	0.39
SF-3	3.7	0.00706	2.8	0.96	0.22
SF-4	4.0	0.00096	2.4	0.95	0.39
SF-5	4.8	0.00082	2.4	0.99	0.42
SF-8	6.1	N/A	N/A	N/A	N/A
SF-9	6.4	0.00027	2.5	0.92	0.66
SF-10	6.8	0.00004	3.1	0.99	1.30
SF-11	6.9	0.00358	1.7	0.89	0.11
SF-12	7.6	0.00090	2.8	0.99	0.45
SF-13	8.0	0.00025	3.1	0.95	0.74
SF-14	8.3	0.00003	2.7	0.88	1.54
SF-15	8.6	0.00002	4.6	0.99	1.41
SF-16	9.3	0.01233	1.1	0.86	0.02
SF-17	10.0	0.00077	2.2	0.77	0.40
SF-19	10.4	0.00409	1.8	0.82	0.13

Table 2-10. Erosion Rate Parameters for 20 to 25 cm Layer

Sediment Core ID	River Mile Location	Proportionality Constant: A	Exponent: n	Correlation Coefficient (R^2)	Critical Shear Stress (Pa)
SF-1	2.4	0.00049	2.8	0.96	0.56
SF-3	3.7	0.00825	2.7	0.98	0.20
SF-4	4.0	0.00056	2.9	0.95	0.55
SF-5	4.8	0.00026	3.0	0.95	0.72
SF-8	6.1	N/A	N/A	N/A	N/A
SF-9	6.4	0.00004	3.2	0.99	1.33
SF-10	6.8	0.00006	2.7	0.97	1.18
SF-11	6.9	N/A	N/A	N/A	N/A
SF-12	7.6	0.00037	3.5	0.99	0.69
SF-13	8.0	0.00011	3.7	0.84	0.97
SF-14	8.3	0.00003	2.8	0.97	1.42
SF-15	8.6	0.00006	3.1	0.99	1.13
SF-16	9.3	0.01254	1.3	0.99	0.02
SF-17	10.0	0.00003	3.9	0.99	1.36
SF-19	10.4	0.00239	2.6	0.72	0.30

Spatial variation, both horizontal and vertical, in the erodibility of Lower Willamette River sediments in cohesive bed areas was evaluated as follows. The first step in this analysis was to calculate average values of the A and n parameters in Equation 2-29 for each of the five depth intervals. For a log-linear relationship (i.e., Equation 2-29), the average exponent (n_{ave}) value for a depth interval is the arithmetic average of the n values for the cores within the interval. The average proportionality constant (A_{ave}) is determined by calculating the log-average value:

$$\log(A_{ave}) = (1/K)\ \sum \log(A_k) \qquad (2\text{-}65)$$

where K is equal to the number of cores (i.e., 15). Using this approach, the average erosion parameters for the five layers in the bed model are listed in Table 2-11.

Table 2-11. Vertical Variation in Average Erosion Rate Parameters

Depth Interval	Average Proportionality Constant: A_{ave}	Average Exponent: n_{ave}	Critical Shear Stress (Pa)
Layer 1: 0 – 5 cm	0.00155	2.2	0.28
Layer 2: 5 – 10 cm	0.00048	2.9	0.58
Layer 3: 10 – 15 cm	0.00052	2.7	0.55
Layer 4: 15 – 20 cm	0.00062	2.6	0.49
Layer 5: 20 – 25 cm	0.00032	2.9	0.66

Vertical variation in the average erosion rate properties for the five depth intervals was quantified using the following procedure. First, calculate the average value of gross erosion rate for depth interval i (${}^{ave}E_{gross,i}$, where i ranges from 1 to 5):

$${}^{ave}E_{gross,i} = 1/N \ \Sigma\, A_{ave,i}\, \tau^{n,ave,i} \tag{2-66}$$

where the summation is over the bed shear stress range of 0.05 to 3 Pa in increments of 0.05 Pa, so N is equal to 60. Values of $A_{ave,i}$ and nave,i for depth interval i are given in Table 2-11. Using the values of ${}^{ave}E_{gross,i}$ for the five depth intervals, the average erosion rate ratios for depth interval i ($R_{ave,i}$) was calculated using:

$$R_{ave,i} = {}^{ave}E_{gross,i} / {}^{ave}E_{gross,1} \tag{2-67}$$

where i ranges from 1 through 5. Thus, $R_{ave,i}$ represents the ratio of the erodibility of depth interval i to the average erodibility of depth interval 1 (i.e., 0 to 5 cm layer); $R_{ave,1}$ is equal to one. The vertical variation in $R_{ave,i}$ is shown on Figure 2-52. These results show that the average erodibility of Lower Willamette River sediment in cohesive bed areas tends to decrease with increasing depth in the bed, which is a typical characteristic of a cohesive sediment bed and is primarily due to increasing consolidation with increasing depth. Quantification of the changes in bed erodibility with depth in the bed (e.g., erodibility of 20 to 25 cm layer is about four times less than the erodibility of 0 to 5 cm layer) aids in interpreting predictions of the sediment transport model, especially during high-flow events.

A similar approach was used to quantify spatial differences in bed erodibility of surface layer (i.e., 0 to 5 cm layer) within the horizontal plane in the Lower Willamette River. The average gross erosion rate for layer 1 (0 to 5 cm layer) in core k was calculated as follows:

$${}^{ave}E_{gross,1,k} = 1/N \ \Sigma\, A_{1,k}\, \tau^{n,1,k} \tag{2-68}$$

where the summation is over the bed shear stress range of 0.05 to 3 Pa in increments of 0.05 Pa, so N is equal to 60. Values of $A_{1,k}$ and $n_{1,k}$ for layer 1 in core k are given in Table 2-6. Using the value of ${}^{ave}E_{gross,1}$ for layer 1, the erosion rate ratio for layer 1 in core k was calculated using:

$$R_{1,k} = {}^{ave}E_{gross,1,k} / {}^{ave}E_{gross,1} \tag{2-69}$$

Thus, $R_{1,k}$ represents the ratio of the erodibility of layer 1 in core k to the average erodibility of layer 1 (0 to 5 cm) sediment. The spatial distribution of $R_{1,k}$ for the 15 cores is presented on Figure 2-53. Nine of the 15 cores are below one (i.e., below the average erodibility of layer 1 sediment). Core SF-11 (RM 6.9) has the lowest erodible sediment in layer 1 (about 25 percent of the average value), with core SF-8 (RM 6.1) having the highest erodibility in layer (about 280 percent greater than the average value).

Sedflume data from 15 cores are not sufficient to use standard interpolation methods to develop a reliable horizontal distribution of erosion properties. In addition, no spatial patterns in the erosion rate ratio of surface-layer sediment are evident on Figure 2-53. Furthermore, no correlation was found between erosion properties and measured bed properties (i.e., dry density, D_{50}, D_{90}, silt/clay content). Thus, developing a credible spatial distribution of erosion parameters in the horizontal plane is problematic. Therefore, it was assumed that the average erosion rate parameters (i.e., A_{ave} and n_{ave} as listed in Table 2-10) for a given depth interval are spatially constant in the horizontal plane within cohesive bed areas. By assuming that the erosion parameters are spatially constant in the horizontal plane, the erosion parameters only vary in the vertical direction. The potential effects of this approximation on model predictions are evaluated during the sensitivity analysis (see Section 2.3.7).

2.3.5 Sediment Transport Model Inputs: Boundary Conditions

The incoming sediment load at the upstream boundary of the Lower Willamette River was estimated using total suspended sediment (TSS) concentration data collected at the USGS Portland gauging station between 1974 and 2009. A sediment rating curve (i.e., correlation between TSS concentration data and flow rate) was developed using the data obtained at the USGS gauging station (Figure 2-54). The Lower Willamette River sediment rating curve has the following behavior: 1) negligible correlation between TSS concentration and flow rate for discharge less than the average flow rate in the Lower Willamette River (33,200 cfs); and 2) positive correlation between TSS concentration and flow rate for discharge greater than the average flow rate. To correct for bias introduced when performing log-linear regressions analysis on the data, the minimum variance unbiased estimator (MVUE) method of Cohn et al. (1992) was used resulting in:

$$C_{LWR} = 9 \quad \text{for } Q_{LWR} < Q_{LWR,ave} \qquad (2\text{-}70)$$

$$= 1.3\,(Q/10{,}000)^{1.65} \quad \text{for } Q \geq Q_{,LWR,ave}$$

Where,
C_{LWR} = TSS concentration (mg/L)
Q_{LWR} = Lower Willamette River flow rate (cfs)
$Q_{LWR,ave}$ = average Lower Willamette River flow rate (33,200 cfs).

For the low-flow regime (i.e., less than average flow rate), C_{LWR} was assumed to be equal to the average value of the TSS concentration data (9 mg/L). The correlation coefficient for the log-linear correlation between TSS concentration and flow rate in the high-flow regime is 0.62.

The composition of the incoming sediment load to the Lower Willamette River was estimated using fine sediment content data (i.e., clay/silt content of suspended sediment load) collected at the USGS Portland gauging station. The relationship between clay/silt content (i.e., class 1 sediment content) and flow rate is shown on Figure 2-55. This relationship indicates the following: 1) negligible correlation between class 1 content and flow rate when flow rate is less than twice the average value; and 2) decreasing class 1 content with increasing flow rate when flow rate is greater than twice the average value. An analysis of these data yielded this functional relationship between class 1 content and flow rate:

$$f_1 = 0.92 \qquad \text{for } Q_{LWR} < 2\, Q_{LWR,ave} \qquad (2\text{-}71)$$

$$= 1.01 - 0.014\,(Q_{LWR}/10{,}000) \qquad \text{for } Q_{LWR} \geq 2\, Q_{LWR,ave}$$

where f_1 is class 1 content in suspended sediment load. The class 2 content in the suspended sediment load (i.e., fine sand content) is given by:

$$f_2 = 1 - f_1 \qquad (2\text{-}72)$$

It was assumed that the class 3 and 4 contents (i.e., medium/coarse sand and gravel) in the incoming sediment load of the Lower Willamette River were zero.

The incoming sediment load at the upstream boundary in the Columbia River was estimated using TSS concentration data collected at the USGS gauging station at Vancouver between 1963 and 1969. The correlation between TSS concentration and flow rate in the Columbia River is shown on Figure 2-56. Based on these data, the sediment rating curve for the Columbia River is:

$$C_{CR} = 0.26\,(Q_{CR}/10{,}000)^{1.53} \qquad (2\text{-}73)$$

Where,
C_{CR} = TSS concentration (mg/L)
Q_{CR} = the Columbia River flow rate (cfs).

The correlation coefficient for the log-linear correlation between TSS concentration and flow rate is 0.13.

The Lower Willamette River sediment rating curve (Equation 2-70) was used to estimate the incoming sediment load to the Lower Willamette River for the 36-year period from 1973 through 2008. Variations in the annual sediment load during this 36-year period are shown on Figure 2-57, with the long-term average being 1,200,000 metric tons/year (MT/yr). For the 7-year period from 2002 through 2008, the average load was 900,000 MT/yr. Seasonal variations in the annual sediment are presented on Figure 2-58, where 73 percent of the annual load occurs during the winter months (December, January, and February).

2.3.6 Sediment Transport Model Calibration Approach and Results

The primary calibration target for the sediment transport model was bed elevation change in the Study Area (RM 1.9 to 11.8) over a multi-year period. The large amount of bed elevation data available made it possible to evaluate the performance of the sediment transport model over a wide range of spatial scales, ranging from the entire Study Area (about 1,920 acres excluding the areas where capping or dredging occurred between surveys) to single grid cells (about 1 acre). The high density of bed elevation data in the Lower Willamette River is unique for a sediment transport model. Typically, less than 10 data points for NSR (based on analysis of radioisotope core data) are available for calibration of a sediment transport model. A relatively high number of NSR data points (about 60 values) were available for evaluating the performance of the Lower Duwamish Waterway sediment transport model (QEA 2008). For the Lower Willamette River sediment transport model, approximately 1,700 values of bed elevation change were used during the model calibration process. Each Lower Willamette River data point represents bed elevation change within a single grid cell, with the average area of a grid cell being 1.1 acres (range of 0.6 to 1.9 acres) within the Site.

Bathymetric surveys were conducted within the Study Area at five different times between 2002 and 2009: 1) January 2002; 2) July/September 2002; 3) May 2003; 4) February 2004; and 5) January 2009. The bathymetry data were analyzed by David Evans and Associates (DEA) to produce bed elevation change data for the four time periods between surveys. Multibeam bathymetry surveys have a typical measurement error range of 0.5 feet resulting in an uncertainty range of 1 foot for bed elevation changes between two surveys. Thus, the uncertainty in bed elevation change decreases as the time period between two surveys increases. Therefore, after a preliminary screening of the data, it was considered more representative to consolidate the first two time periods and analyze the bed elevation changes for the following time periods: 1) January 2002 to May 2003 (16 months); 2) May 2003 to February 2004 (10 months); and 3) February 2004 to January 2009 (58 months). An evaluation of the bed elevation change data during these three time periods was conducted to determine what data would be most useful for calibrating the sediment transport model.

The average incoming sediment load to the Lower Willamette River for the 7-year period (January 2002 to January 2009) was 944,000 MT/yr. A summary of the incoming sediment load for the three time periods between 2002 and 2009 when bed elevation change are available is given in Table 2-12. The temporal distribution of the total upstream load between the three time periods is presented on Figure 2-59, where 72 percent of the load occurred during February 2004 to January 2009.

Table 2-12. Incoming Sediment Load to Lower Willamette River Between 2002 and 2009

Time Period	Length of Time Period (months)	Annual Sediment Load (MT/yr)	Total Sediment Load During Period (MT)
January 2002 to May 2003	16	915,000	1,220,000
May 2003 to February 2004	10	809,000	674,000
February 2004 to January 2009	58	972,000	4,698,000
January 2002 to January 2009	84	937,000	6,592,000

Notes:
MT = metric tons
MT/yr = metric tons per year

The bed elevation change data were analyzed and used to construct sediment mass balances for the 9.9-mile reach of the Lower Willamette River extending from RM 1.9 to 11.8 during the 7-year period between January 2002 and January 2009 (see Figures 2-60 through 2-63). The total mass of sediment eroded or deposited within the 9.9-mile reach of the Lower Willamette River was calculated using the bed elevation change data for a particular time period. The average NSR for the entire 9.9-mile reach was also estimated from the bed elevation change data. This information, combined with the upstream sediment load, was used to construct a sediment mass balance for a specific time period and calculate the trapping efficiency of the 9.9-mile reach, where trapping efficiency is the portion of the upstream sediment load deposited in the reach. A summary of the mass balance results is given in Table 2-13. The temporal distribution of net deposition between the three time periods is presented on Figure 2-64, where 92 percent of the total net deposition occurred during February 2004 to January 2009. Variation in the average NSR during the different time periods between January 2002 and January 2009 is presented on Figure 2-65.

Table 2-13. Sediment Mass Balances for RM 1.9 to 11.8 Region for 2002 to 2009 Period

Time Period	Net Deposition Rate (MT/yr)	Average Net Sedimentation Rate (cm/yr)	Trapping Efficiency (%)
January 2002 to May 2003	-1,400 (net erosion)	0.0	0
May 2003 to February 2004	154,000	2.1	18
February 2004 to January 2009	262,000	3.5	26
January 2002 to January 2009	197,000	2.6	20

Notes:
cm/yr = centimeters per year
MT/yr = metric tons per year

The results of this analysis of the bed elevation change data have the following implications for calibration of the sediment transport model. Net deposition processes in the 9.9-mile reach (RM 1.9 to 11.8) during the 16-month period from January 2002 to May 2003 are significantly different than during the period from May 2003 to January 2009, even though the incoming sediment load is similar. Thus, the 16-month period (January 2002 to May 2003) appears to be anomalous and using bed elevation change data collected during that period for calibration would have produced a sediment transport model that would predict unrealistically low NSRs. Therefore, the 68-month period from May 2003 to January 2009 was selected as the model calibration period because it will produce a robust model that is representative of long-term sediment transport processes in the Study Area (RM 1.9 to 11.8).

The sediment transport model was calibrated using bed elevation change data collected during the 68-month period from May 2003 to January 2009. The hydrograph in the Lower Willamette River for this time period is shown on Figure 2-66. Several high-flow events (i.e., greater than 100,000 cfs) occurred during the calibration period, with the maximum flow rate happening during a 2-year flood in January 2006. Three input parameters were adjusted during the calibration process: 1) effective diameter of class 2 (fine sand) sediment; 2) effective diameter of class 3 (medium/coarse sand) sediment; and 3) exponent (N) in active layer thickness formulation for non-cohesive bed (Equation 2-59). The data-based and calibrated effective diameters for all sediment classes are listed in Table 2-14. The exponent (N) in the active layer thickness equation for the non-cohesive bed was set at a value of 0.5.

Table 2-14. Characteristics of Sediment Particle Size Classes

Sediment Size Class	Particle Size Range (μm)	Data-Based Effective Particle Diameter (μm)	Calibration Effective Particle Diameter (μm)
1: clay and silt	< 62	Not Applicable	Not Applicable
2: fine sand	62 – 250	140	100
3: medium, coarse sand	250 – 2,000	550	700
4: gravel	> 2,000	2,800	2,800

Notes:
μm = micrometer

Model performance was evaluated over a range of spatial scales within the Site (RM 1.9 to 11.8). The calibration results will be examined starting with the largest spatial scale (i.e., entire Study Area, about 1,920 acres) and then proceed to smaller spatial scales, with the small scale being a single grid cell (about 1 acre). The sediment mass balance predicted by the model within the Study Area (RM 1.9 to 11.8) for the 68-month calibration period is presented on Figure 2-67. The model predicted a net deposition rate of 203,000 MT/yr for the entire Site, which is within 6 percent of the data-based estimate (192,000 MT/yr). If regions where capping and dredging activities occurred during the calibration period are excluded from the comparison, then the predicted and measured net deposition rate is reduced by 14,000 MT/yr and 8,000 MT/yr, respectively. The data-based and predicted trapping efficiencies for the Site during the calibration period are 20 percent and 21 percent, respectively. Comparison of measured (data-based) and predicted regions of net erosion and deposition within the Site during the 68-month period are shown on Figure 2-68. These results show that for the largest spatial scale (entire Study Area), the model adequately predicts the net deposition mass and trapping efficiency with a slight tendency to over-predict sedimentation. In addition, the overall spatial distribution of erosion and deposition areas is well represented by the model.

Spatial distributions of laterally averaged (i.e., cross-channel average) bed elevation change predicted by the model during the 68-month calibration period were compared to measured values over three spatial scales: 1) 1-mile average (about 200 acres); 2) 0.5-mile average (about 100 acres); and 3) 1/8-mile average (about 22 acres). The 1/8-mile laterally averaged bed elevation changes correspond to one row of grid cells across the river channel. The predicted and measured longitudinal (along channel) distributions of laterally-averaged values for the 1-mile, 0.5-mile and 1/8-mile averages are shown on Figures 2-69, 2-70 and 2-71, respectively. The model-data comparisons on these figures demonstrate that the model adequately simulates longitudinal variations in laterally-averaged bed elevation change from 1-mile to 1/8-mile spatial scales.

A spatial scale analysis (QEA 2008) was conducted to determine the relationship between model predictive capability and spatial scale within the Study Area (RM 1.9 to 11.8). Absolute differences between predicted and measured bed elevation changes were calculated for different zones located in the RM 1.9 to 11.8 region, with zonal areas ranging from about 1 acre (i.e., a single grid cell) to about 1,920 acres (entire RM 1.9-11.8 region). The zonal analysis separated the Site into zones based on rows of grid cells across the Lower Willamette River channel; on average, there are 25 grid cells in the cross-channel direction. An idealized channel consisting of 11 rows in the longitudinal direction, and 5 grid cells in the cross-channel direction is used to illustrate this approach. The smallest zones in this analysis are 1-row zones, which results in 11 zones for the idealized channel (Figure 2-72). The next-largest zones are composed of 2-row zones (Figure 2-73). For the 2-row zones, each zone is shifted by one row with respect to location, so that adjacent zones overlap by one row. This process results in ten 2-row zones. Similarly, there are nine 3-row zones, eight 4-row zones, and, finally, two 10-row zones (Figure 2-74).

The primary results of the spatial-scale analysis are shown in Figure 2-75. This figure presents the absolute difference for zonal spatial-scales ranging from about 1 to 1,920 acres. The solid dots in Figure 2-75 represent the average absolute difference, with the 95 percent confidence interval about the average shown as error bars. The following general conclusions were derived from the spatial-scale analysis: 1) the average absolute difference is less than +/- 0.1 centimeters per year (cm/yr) for all spatial scales, indicating that the model is not biased and has approximately the same predictive capability, on average, over the entire range of spatial scales; and 2) for all spatial scales the confidence interval of the model prediction is below $\pm$0.5 cm/yr, which indicates that the model does not tend to significantly over-predict sedimentation rate.

The smallest spatial scale at which the predictive capability of the model was evaluated is a single grid cell. The first step in this evaluation was determining qualitative agreement between erosion and deposition areas (e.g., if the model predicts net deposition in a specific grid cell, is the prediction consistent with the data-based bed elevation change?). The spatial distribution of qualitative agreement between areas of net erosion and deposition during the 68-month calibration period is shown on Figure 2-76. Overall, model predictions of net erosion and deposition on a grid-cell basis agreed with the measured bed elevation change for 75 percent of the grid cells within the Study Area (RM 1.9 to 11.8), which contains a total of 1,909 grid cells (Figure 2-77). Results of a quantitative analysis of model predictions at the grid-cell spatial scale within areas of net erosion and deposition are shown on Figure 2-78 and summarized in Table 2-15. Within the Site, the model predicted net deposition to occur over 88 percent of the total area (1,690 acres) and net erosion over the remaining 12 percent (230 acres) during the time period between May 2003 and January 2009. Thus, these results indicate that significant variability exists in the predictive capability of the sediment transport model at the grid-cell spatial scale.

Table 2-15. Sediment Transport Model Performance at Grid-Cell Spatial Scale

Difference Between Predicted and Measured Bed Elevation Change (cm/yr)	Relative Portion of Grid Cells With Predicted Net Erosion (%)	Relative Portion of Grid Cells With Predicted Net Deposition (%)
Less than ±2.5	31	58
±2.5 to ±5	27	26
±5 to ±10	30	12
Greater than ±10	12	4

Notes:
cm/yr = centimeters per year

In summary, the predictive capability of the sediment transport model was evaluated over a wide range of spatial scales, from a single grid cell (about 1 acre) to the entire Study Area (about 1,900 acres), during the calibration process. Extensive analysis of model results demonstrated that:

- Within the Study Area (RM 1.9 to 11.8), the model is able to adequately simulate: 1) large-scale deposition and erosion processes; and 2) longitudinal variations in laterally averaged bed elevation change.
- At grid-cell spatial scales (about 1 acre), model predictions have approximately zero bias on average. However, significant variability exists in model predictive capability at this spatial scale.

Therefore, it is concluded that the Lower Willamette River sediment transport model was successfully calibrated.

As described above, the sediment transport model was calibrated using bed elevation change data in the Study Area. The calibration process, however, did not consider the behavior of sediments that are too fine to settle onto the sediment bed (i.e., wash load). Although wash load by definition does not impact erosion and deposition processes, it has significance in the fate and transport of contaminants as it affects the partitioning of contaminants in the water column. Therefore, existing TSS concentration data collected during the calibration period (Integral et al. 2011) were used to estimate the wash load concentration. During low-flow conditions, it was determined that the wash load concentration is about 5 mg/L. To represent the wash load in the model, an additional sediment class with zero settling speed and a constant concentration of 5 mg/L was added. Only predicted values of TSS concentration are affected by this modification to the model.

2.3.7 Evaluation of Model Reliability

The calibrated sediment transport model is a reliable tool for erosion, deposition, and transport processes over multi-year periods in the Lower Willamette River. However, uncertainty exists in model predictions because of uncertainty in model inputs or parameters. The effects of input uncertainty on model predictions were evaluated through a sensitivity analysis.

Based on a similar uncertainty analysis conducted for a sediment transport model of the Lower Duwamish Waterway, which is an estuary near Seattle, Washington (QEA 2008), the effects of five inputs on model uncertainty were chosen for this analysis: 1) upstream sediment load in Lower Willamette River; 2) upstream sediment load in Columbia River: 3) composition of upstream sediment load in Lower Willamette River; 4) erosion rate parameters for cohesive sediment; and 5) effective bed roughness (D_{90}). Lower- and upper-bound limits of these there inputs were specified as follows:

- *Upstream sediment load in Lower Willamette River:* ± a factor-of-two with respect to base-case simulation
- *Upstream sediment load in Columbia River:* ± a factor-of-two with respect to base-case simulation
- *Composition of upstream sediment load in Lower Willamette River:* ± 5 percent class 1 content with respect to base-case simulation
- *Erosion rate parameters for cohesive sediment:* lower-bound corresponds to least erodible core and upper-bound corresponds to most erodible core
- *Effective bed roughness:* mean $\pm$ 2 standard errors

Table 2-16 presents additional information regarding the parameters used for the sensitivity analysis.

Table 2-16. Summary of Parameters Varied in Sensitivity Analysis

Parameters	Base Case	Lower Bound	Upper Bound
Average upstream sediment load in Lower Willamette River (MT/yr)	951,000	475,000	1,901,000
Average upstream sediment load in Columbia River (MT/yr)	3,769,000	1,884,000	7,537,000
Composition of upstream sediment load in Lower Willamette River f_1: fraction of class 1 Q: Discharge Q_m: Mean discharge = 33,200 cfs	if $Q < 2Q_m$, $f_1 = 0.92$ else, $f_1 = 1.01–0.014\frac{Q}{10000}$	if $Q < 2Q_m$, $f_1 = 0.87$ else, $f_1= 0.96-0.014\frac{Q}{10000}$	if $Q < 2Q_m$, $f_1 = 0.97$ else, $f_1= 1.06-0.014\frac{Q}{10000}$
Erosion rate parameters for cohesive sediment	Average Sedflume core See Table 2-11	Least erodible Sedflume core (SF-11) See Tables 2-6 to 2-10	Most erodible Sedlume core (SF-8) See Tables 2-6 to 2-10
Effective bed roughness: D_{90} (µm)	Spatial variable D_{90}	Cohesive bed = 286 µ, Non cohesive bed = 1196 µ	Cohesive bed = 327 µ, Non cohesive bed = 1355 µ

To evaluate the effects of possible interactions between the three inputs, a factorial analysis was conducted, which resulted in 32 simulations to account for all of the possible combinations of the bounding limits of the five inputs. The parameter sets used in the 32 uncertainty simulations are provided in Table 2-17, where "lower" refers to lower-bound value and "upper" refers to upper-bound value.

Table 2-17. Bounding Limits for Sensitivity Simulations

Simulation Number	Lower Willamette River Sediment Load Bound	Columbia River Sediment Load Bound	Lower Willamette River Sediment Load Composition Bound	Erosion Rate Parameters Bound	Effective Bed Roughness Bound
1	Lower	Lower	Lower	Lower	Lower
2	Lower	Lower	Lower	Lower	Upper
3	Lower	Lower	Lower	Upper	Lower
4	Lower	Lower	Lower	Upper	Upper
5	Lower	Lower	Upper	Lower	Lower
6	Lower	Lower	Upper	Lower	Upper
7	Lower	Lower	Upper	Upper	Lower
8	Lower	Lower	Upper	Upper	Upper
9	Lower	Upper	Lower	Lower	Lower
10	Lower	Upper	Lower	Lower	Upper
11	Lower	Upper	Lower	Upper	Lower
12	Lower	Upper	Lower	Upper	Upper
13	Lower	Upper	Upper	Lower	Lower
14	Lower	Upper	Upper	Lower	Upper
15	Lower	Upper	Upper	Upper	Lower
16	Lower	Upper	Upper	Upper	Upper
17	Upper	Lower	Lower	Lower	Lower
18	Upper	Lower	Lower	Lower	Upper
19	Upper	Lower	Lower	Upper	Lower
20	Upper	Lower	Lower	Upper	Upper
21	Upper	Lower	Upper	Lower	Lower
22	Upper	Lower	Upper	Lower	Upper
23	Upper	Lower	Upper	Upper	Lower
24	Upper	Lower	Upper	Upper	Upper
25	Upper	Upper	Lower	Lower	Lower
26	Upper	Upper	Lower	Lower	Upper
27	Upper	Upper	Lower	Upper	Lower
28	Upper	Upper	Lower	Upper	Upper
29	Upper	Upper	Upper	Lower	Lower
30	Upper	Upper	Upper	Lower	Upper
31	Upper	Upper	Upper	Upper	Lower
32	Upper	Upper	Upper	Upper	Upper

The sensitivity simulation results were analyzed, with simulations 11 and 26 producing the ultimate lower- and upper-bound results, respectively. Further analysis of the sensitivity results showed that simulations 7 and 12 yielded results which correspond to realistic lower- and upper-bound predictions (i.e., within a factor of two of the base case simulation). The realistic lower- and upper-bound simulations (i.e., parameters sets for simulations 7 and 12) were used as input for the contaminant fate and transport model to evaluate the impact of uncertainty in the sediment transport model inputs on the contaminant fate and transport model (see Appendix Ha for more discussion on the contaminant fate model sensitivity analysis). Base case simulation results are compared to lower- and upper-bound results for the realistic and ultimate model parameter sets on Figure 2-79.

3.0 DIAGNOSTIC AND PROGNOSTIC MODELING ANALYSES

3.1 SEDIMENT BED STABILITY DURING HIGH-FLOW EVENTS

The effects of high-flow events on bed stability in the Lower Willamette River were evaluated by using the sediment transport model to simulate the February 1996 flood. The peak flow rate during the 1996 flood was 420,000 cfs, which likely represents a high-flow event with a return period of at least 100 years. Therefore, the bed stability analysis was conducted using flow conditions that are representative of an extreme high-flow event.

The predicted sediment mass balance for the Study Area (RM 1.9 to 11.8) during the 1996 flood is shown on Figure 3-2. The incoming sediment load was 1,260,000 MT, which was estimated using Equation 2-70. This incoming load, which occurred during a 14-day period, is about 33 percent greater than the average annual incoming load during the 68-month calibration period (i.e., 950,000 MT/yr). Overall, net erosion occurred within the Study Area, with 182,000 MT of sediment being predicted to be eroded and transported out of the Study Area. This amount of sediment is approximately equal to the amount of net deposition that occurred on an annual basis during the calibration period (see Figure 2-67).

The spatial distribution of bed elevation change that was predicted during the 1996 flood is shown on Figure 3-3. Overall, net erosion occurred in 54 percent of the bed area within the RM 1.9 to 11.8 region. Predicted maximum scour depths during 1996 flood are presented on Figure 3-4. The relative areal distribution of various ranges of scour depths is tabulated in Table 3-1. The maximum scour depth was predicted to be 62 cm, which occurred near RM 5.3. This information was used in evaluations of the potential for rare high-flow events to expose elevated contaminant concentrations buried at depth in the bed, as discussed in Section 5.7 of the main text.

Table 3-1. Relative Areal Distribution of Maximum Scour Depths During 1996 Flood

Maximum Scour Depth (cm)	Relative Portion of Area Within River Mile 1.9 to 11.8 Region (%)
0 to 10	43
10 to 20	13
20 to 30	16
Greater than 30	1
Net Depositional	27

Notes:
cm = centimeter

A diagnostic analysis of the 1996 flood simulation was conducted to develop insights about model performance at specific locations within the Study Area. A total of 15 locations were selected for the diagnostic analysis, with 10 locations in cohesive bed areas and 5 locations in non-cohesive bed areas (Figure 3-5). The locations were selected such that a range of sediment transport regimes were represented (i.e., different rates of net sedimentation or net erosion), see Table 3-2. The NSRs in Table 3-2 correspond to predicted values for the 68-month calibration simulation; negative values in the table represent net erosion.

Table 3-2. Locations of Grid Cells Used in Diagnostic Analysis of 1996 Flood

Location Label	River Mile Location	Type of Bed	Net Sedimentation Rate (cm/yr)
C-1	3.22	Cohesive	0.0
C-2	3.28	Cohesive	-0.19
C-3	3.64	Cohesive	0.08
C-4	4.30	Cohesive	1.3
C-5	4.40	Cohesive	5.0
C-6	6.81	Cohesive	-0.21
C-7	7.19	Cohesive	0.02
C-8	7.42	Cohesive	0.2
C-9	7.89	Cohesive	24
C-10	10.82	Cohesive	3.8
NC-1	5.29	Non-cohesive	0.76
NC-2	5.66	Non-cohesive	-0.7
NC-3	8.93	Non-cohesive	2.1
NC-4	9.22	Non-cohesive	1.5
NC-5	10.31	Non-cohesive	13

Notes:
cm/yr = centimeters per year

The diagnostic analysis of the 1996 flood simulation examined temporal variations in the following predictions at specific grid cell locations: depth-averaged current velocity; bed shear stress; TSS concentrations (total, classes 1, 2, 3, 4); bed elevation change; erosion flux; deposition flux; and surface-layer bed content of classes 1, 2, 3, and 4. The results of the diagnostic analysis at the 15 grid cell locations are presented on Figures 3-6 through 3-35.

Eight of the ten cohesive bed locations were predicted to be net erosional during the 1996 flood, with net erosion depths ranging between about 10 and 25 cm and peak bed shear stresses greater than 1.5 Pa. The two cohesive bed locations that were net depositional during the 1996 flood (i.e., locations C-5 and C-9) are the two locations where the long-term NSR was predicted to be greater than 4 cm/yr. Peak bed shear stresses were less than about 1 Pa at the cohesive bed locations that were net depositional.

At the non-cohesive bed locations, three locations were predicted to be net erosional during the 1996 flood, with maximum net erosion depths of about 5 cm and peak bed shear stresses between 4 and 6 Pa. The two non-cohesive bed locations that were net depositional during the 1996 flood (i.e., locations NC-3 and NC-5) are locations where the long-term NSR was greater than 2 cm/yr. Peak bed shear stresses were less than about 1 Pa at the non-cohesive bed locations that were net depositional.

3.2 LONG-TERM SEDIMENT TRANSPORT PROCESSES

Sediment transport in the Lower Willamette River was simulated over a 45-year period. This simulation was conducted to: 1) evaluate the behavior of the sediment transport model over a multi-year period; and 2) provide coupling output for the contaminant fate and transport model such that the effects of remedial alternatives could be evaluated over a 45-year period (see Appendix Ha). The 45-year sediment transport simulation was constructed as follows: 1) simulate 30-year period (1979 through 2008) with the hydrodynamic model; and 2) create hydrodynamic model inputs for the sediment transport model by repeating the first 15 years (i.e., 1979 through 1993) and using those inputs for years 31 through 45 of the 45-year sediment transport simulation. This approach produces a reasonable approximation of long-term hydrodynamic conditions in the Study Area. Temporal variations in the annual average flow rate in the Lower Willamette River for 1979 through 2008 are displayed on Figure 3-36, with the lowest and highest values occurring during 1992 and 1996 respectively. The average annual incoming sediment load to the Lower Willamette River during this 30-year period was about 1,200,000 MT/yr (see Figure 2-57).

The 45-year simulation was conducted using "base case" model inputs (i.e., calibration parameters, upstream sediment load based on Equation 2-70). The only input that needed to be modified for the 45-year simulation was the downstream boundary in the hydrodynamic model due to lack of WSE data at the NOAA gauge located at St. Helens before 2002. Thus, water surface elevations predicted by NOAA using tidal harmonics for the period between 1979 and 2001 were used to specify the downstream boundary condition. Predicted WSE will tend to be lower than actual values because changes in mean WSE due to changes in upstream river discharge are not incorporated into the NOAA tidal harmonics model. Lower mean WSE at the downstream boundary will produce an over-prediction of current velocities by the hydrodynamic model. The over-prediction of current velocities will tend to decrease the deposition rates predicted by the sediment transport model and, thus, the contaminant fate and transport model will generate more conservative predictions (i.e., slower rate of natural recovery will be predicted).

Sensitivity of the long-term simulations to uncertainty in model input parameters was evaluated by conducting 30-year simulations using the realistic lower- and upper-bound parameter sets discussed in Section 2.3.7 (i.e., sensitivity simulations 7 and 12). Although the full 45-year period was simulated to provide the necessary inputs for the contaminant fate and transport model, the two sensitivity simulations were conducted for only the first 30 years of the 45-year period. The predicted sediment mass balance for the base-case 30-year (1979 to 2008) simulation is shown on Figure 3-37. The Study Area (RM 1.9 to 11.8) is net depositional, with a trapping efficiency of 19 percent. Sediment mass balances for the lower- and upper-bound 30-year simulations are presented on Figures 3-38 and 3-39, respectively. The trapping efficiencies for the bounding simulations are approximately equal to the trapping efficiency for the base-case simulation. This result indicates that variations in the five model parameters/inputs used in the sensitivity analysis do not affect the large-scale behavior of the model. The bounding simulations also demonstrate that the net deposition rate in the Study Area is approximately linearly correlated to the incoming sediment load in the Lower Willamette River (i.e., net deposition rate increases by 100 percent when the incoming sediment load is increased by 100 percent).

The relative portions of the Study Area with net erosion and deposition during the 30-year base-case simulation are shown on Figure 3-40. Similar graphical displays for the lower- and upper-bound simulations are presented on Figure 3-41 and 3-42, respectively. Numerical comparisons of the base-case, lower- and upper-bound simulations are provided in Table 3-3. Longitudinal (along channel) variations in laterally-averaged bed elevation change for the base-case, lower- and upper-bound simulations are presented on Figure 3-43.

Table 3-3. Relative Portions of Study Area with Net Erosion/Deposition, 30-Year Simulation

Range of Net Erosion and Deposition	Portion of the Study Area: Base-Case Simulation (%)	Portion of the Study Area: Lower-Bound Simulation (%)	Portion of the Study Area: Upper-Bound Simulation (%)
Erosion: < -5 cm/yr	1	1	0
Erosion: 0 to -5 cm/yr	13	14	6
Deposition: 0 to 5 cm/yr	56	74	49
Deposition: > 5 cm/yr	30	11	45

Notes:
cm/yr = centimeters per year

Model predictions for the 30-year simulation were evaluated at the same 15 locations used in the diagnostic analysis of the 1996 flood simulation (see Table 3-2 and Figure 3-5 for locations). The analysis of the 30-year simulation examined temporal variations in the following model predictions at specific grid cell locations: 1) bed elevation change; and 2) surface-layer bed content of classes 1, 2, 3, and 4. Results of the 30-year diagnostic analysis at the 15 grid cell locations are presented on Figures 3-44 through 3-58.

Eight of the ten cohesive bed locations were predicted to be net depositional during the 30-year period, with NSRs typically ranging between about 0.1 and 33 cm/yr. The two cohesive bed locations that were net erosional during the 30-year period (i.e., locations C-1 and C-6) had average net erosion rates ranging between about -0.1 and -1.3 cm/yr. At the five non-cohesive bed locations, three locations were predicted to be net depositional during the 30-year period, with NSRs ranging between about 1.5 and 15 cm/yr. The other two locations, NC-1 and NC-2, presented average net erosion rates of -1.2 and -3.4 cm/yr, respectively. The effects of the 1996 flood on bed elevation change are clearly evident at all 15 locations.

4.0 REFERENCES

Berger, C., M. McKillip, R. Annear, S.J. Khan, and S. Wells, 2004. "Willamette River Basin Temperature TMDL Model: Model Calibration". *Technical Report EWR-02-04. School of Engineering and Applied Sciences, Department of Civil and Environmental Engineering, Portland State University.*

Blumberg, A. F., and Mellor, G. L., 1987. "A description of a three-dimensional coastal ocean circulation model." *Three-dimensional coastal ocean models. Coastal and estuarine sciences: Volume 4,* N. Heaps, ed., American Geophysical Union, Washington, D.C., 1-16.

Burban, P.Y., Y.J. Xu, J. McNeil, and W. Lick, 1990. "Settling speeds of flocs in fresh water and seawater." *J. Geophys. Res.*, 95(C10):18,213-18,220.

Cheng, N.S., 1997. Simplified settling velocity formula for sediment particle. *ASCE J Hydr Engr.* 123(2):149-152.

Cohn, T.A., D.L. Caulder, E.J. Gilroy, L.D. Zynjuk, and R.M. Summers, 1992. The validity of a simple statistical model for estimating fluvial constituent loads: an empirical study involving nutrient loads entering Chesapeake Bay. *Water Resour. Res*., 28(9):2353-2363.

DEA (David Evans and Associates), 2002. Lower Willamette River Multibeam Bathymetric Survey Report, April 2002. Prepared for Striplin Environmental Associates, Olympia, WA.

Garcia, M. and G. Parker, 1991. Entrainment of bed sediment into suspension. *ASCE J Hydr Engr,* 117(4):414-435.

Gessler, J., 1967. *The beginning of bedload movement of mixtures investigated as natural armoring in channels.* W.M. Keck Laboratory of Hydraulics and Water Resources, California Institute of Technology, Translation T-5.

GeoSea, 2001. A Sediment Trend Analysis STA® on the Lower Willamette River, April 2001.

Guy, H.P., D.B. Simons, and E.V. Richardson, 1966. *Summary of alluvial channel data from flume experiments, 1956-1961.* U.S. Geological Survey Professional Paper 462-I. Washington, D.C.

Hamrick, J.M., 1992. *A Three-Dimensional Environmental Fluid Dynamics Code: Theoretical and Computational Aspects.* Special Report 317 in Applied Marine Science and Ocean Engineering, Virginia Institute of Marine Science, VA.

Helsel, D.R. and R.M. Hirsch, 2002. Statistical Methods in Water Resources. Chapter A3, Book 4, Hydrologic Analysis and Interpretation, Techniques of Water-Resources Investigations of the United States Geological Survey [Online]. USGS, Washington, DC. Updated 2002.

Integral Consulting, Inc. 2005a. Portland Harbor RI/FS Round 2 Surface and Beach Sediment Field Sampling Report (Draft). Prepared for Lower Willamette Group.

Integral Consulting, Inc. 2005b. Portland Harbor RI/FS Round 2A Surface Water Field Sampling Report (Draft). Prepared for Lower Willamette Group.

Integral Consulting, Inc. 2006. Portland Harbor RI/FS Round 2 Hydrodynamic/Sediment Transport Modeling Data Collection Field Sampling Report (Draft). Prepared for Lower Willamette Group.

Integral Consulting Inc., Windward Environmental LLC, Kennedy/Jenks Consultants, Anchor QEA, LLC. 2011. Portland Harbor RI/FS Remedial Investigation Report. Draft Final. Prepared for the Lower Willamette Group. Portland, OR.

IACWD (Interagency Advisory Committee on Water Data), 1982. Guidelines for Determining Flood Flow Frequency. Bulletin 17B of the Hydrology Subcommittee. USGS. Reston, VA.

Jepsen R, J. Roberts, and W. Lick, 1997. Effects of bulk density on sediment erosion rates. *Water Air Soil Pollution*, 99:21-31.

Jones, C.A. and W. Lick, 2001. *SEDZLJ: A Sediment Transport Model.* Final report. University of California, Santa Barbara, California. May 29, 2001.

Karim, M.F. and J.F. Kennedy, 1981. *Computer-based predictors for sediment discharge and friction factor of alluvial streams, IIHR report no. 242.* Univ. of Iowa, Iowa City, Iowa.

Parker, G., 2004. *1D Sediment Transport Morphodynamics, with Applications to Rivers and Turbidity Currents.* E-book located at: http://cee.uiuc.edu/people/parkerg/morphodynamics_e-book.htm.

QEA, 2008. *Lower Duwamish Waterway Sediment Transport Modeling Report, Draft Final.* Prepared for USEPA, Region 10 and Washington State Dept. of Ecology. June 2008.

Rahuel, J.L., F.M. Holly, J.P. Chollet, P.J. Belleudy, and G. Yang, 1989. Modeling of riverbed evolution for bedload sediment mixtures. *ASCE J Hydr Engr*, 115(11):1521-1542.

Roberts J., R. Jepsen, and W. Lick, 1998. Effects of particle size and bulk density on the erosion of quartz particles. *ASCE J Hydr Engr* 124(12):1261-1267.

Sea Engineering, Inc., 2006. Sedflume Analysis, Willamette River, Portland, OR. Prepared for Integral Consulting, Inc.

van den Berg, J.H. and A. van Gelder, 1993. Prediction of suspended bed material transport in flows over silt and very fine sand. *Water Resour. Res*., 29(5):1392-1404.

van Rijn, L.C., 1984a. Sediment transport, part I: bed load transport. *ASCE J Hydr Engr.* 110(10):1431-1456.

van Rijn, L.C., 1984b. Sediment transport, part II: suspended load transport. *ASCE J Hydr Engr.* 110(11):1612-1638.

van Rijn, L.C., 1984c. Sediment transport, part III: bed forms and alluvial roughness. *ASCE J Hydr Engr.* 110(12):1732-1754.

van Rijn, L.C., H. van Rossum, and P. Termes, 1990. Field verification of 2-D and 3-D suspended-sediment models. *ASCE J Hydr Engr*, 116(10):1270-1288.

van Rijn, L.C., 1993. *Principles of Sediment Transport in Rivers, Estuaries and Coastal Seas*. Aqua Publications. Delft, The Netherlands.

Voogt, L., L.C. van Rijn, and J.H. van den Berg, 1991. Sediment transport of fine sands at high velocities. *ASCE J Hydr Engr,* 117(7):869-890.

WEST Consultants, 2004. Hydrodynamics/Sedimentation Modeling for Lower Willamette River, Development of Modeling Approach. Technical Memorandum 001. Prepared for Lower Willamette Group, Portland, OR.

WEST Consultants, 2005a. Phase 1 Results: Hydrodynamic Sedimentation Modeling for Lower Willamette River. Draft Report. Prepared for Lower Willamette Group, Portland, OR.

WEST Consultants, 2005b. Revised Phase 1 Results: Hydrodynamic Sedimentation Modeling for Lower Willamette River. Draft Report. Prepared for Lower Willamette Group, Portland, OR.

WEST Consultants, 2006. Phase 2 Recalibration Results: Hydrodynamic Sedimentation Modeling for Lower Willamette River. Draft Report. Prepared for Lower Willamette Group, Portland, OR.

WEST Consultants and TetraTech, 2009. Revised Phase 2 Recalibration Results: Hydrodynamic Sedimentation Modeling for Lower Willamette River. Draft Report. Prepared for Lower Willamette Group, Portland, OR.

Wright, S., and G. Parker, 2004. Flow resistance and suspended load in sand-bed rivers: simplified stratification model. *ASCE J. Hydr. Engr.* 130(8):796-805.

Ziegler, C.K. and W. Lick, 1988. The transport of fine-grained sediments in shallow waters. *Environmental Geology and Water Sciences* 11:123-132.

Ziegler, C.K, P.H. Israelsson, and J.P. Connolly, 2000. Modeling sediment transport dynamics in Thompson Island Pool, Upper Hudson River. *Water Quality and Ecosystem Modeling* 1:193-222.

Ziegler, C. K. and B. Nisbet, 1994. Fine grained sediment transport in Pawtuxet River, Rhode Island. *ASCE J Hydr Engr,* 120(5):561-576.